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Variations in the Abundance of O^{18} in Ice and Snow from the Kaskawulsh Glacier

by

Donald Stuart Macpherson

A Thesis Submitted to the Faculty of Graduate Studies in Partial Fulfilment of the Requirements for the Degree of Master of Science.

Department of Geology

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UNIVERSITY OF ALBERTA

FACULTY OF GRADUATE STUDIES

The undersigned certify that they have read, and recommend to the Faculty of Graduate Studies for acceptance, a thesis entitled "Variations in the Abundance of O^{18} in Ice and Snow from the Kaskawulsh Glacier", submitted by Donald S. Macpherson, B.Sc., in partial fulfilment of the requirements for the degree of Master of Science.

ABSTRACT

100 samples for this study were taken during the summer of 1963 from the Hubbard and Kaskawulsh glacier systems with the idea of (i) correlating variations in the O^{18}/O^{16} ratio with annual meteorological trends and (ii) using the O^{18}/O^{16} ratio as a natural tracer of glacier flow.

The values found for snow and ice ranged from $\delta = -17.7$ to $\delta = -29.2$, with respect to "Standard Mean Ocean Water." These values compare favourably with the range of O^{18}/O^{16} ratios found by Epstein *et al* (1959) on the Malaspina glacier in the same general vicinity. There exists a definite correlation between snow samples from pit studies on the Kaskawulsh glacier and the glaciology and annual precipitation pattern of the area. It has been demonstrated in this study that subsequent downward percolation of summer melt water produced a homogenization of the O^{18}/O^{16} ratio, and diluted the significant annual pattern noticed prior to the onset of summer melt action. An altitude effect was noticed in the hard ice below the firn line on the Kaskawulsh glacier that tends to support existing theories as to the mode of glacier flow.

ACKNOWLEDGEMENTS

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1. Introduction and General Theory ~~of~~ O^{18}/O^{16} Variations and Their Possible Application to this Problem.

The relative abundance of oxygen isotopes from natural sources obviously depends on the nature of the source. The greatest extremes in O^{18} abundance have been found to be between south polar snow and atmospheric CO_2. Such differences enabled earlier workers to detect isotopic variations by physical methods such as density or specific gravity variations. This required density to be measured to 0.1 to 1.0 parts per million, but was nevertheless done by a number of workers, notably Riesenfield and Chang in 1936. In 1936 Dole first pointed out that specific gravity variations were due principally to variations in the O^{18} content of such waters. Almost immediately specific gravity measurements were applied to snow and ice. Accurate determinations of the O^{18} variations by mass spectrometric means have been made since 1950. A theoretical basis for such investigations has been made through calculations of isotopic variations by Urey (1947). Sharp (1960) has interpreted the significance of O^{18} variations in ice and snow as applied to glaciology.

The abundance of the stable oxygen isotopes in atmospheric oxygen is:

O^{16} - 99.754 atomic %

O^{17} - .0374 atomic %

O^{18} - .2039 atomic %

A discussion of variations in these abundances can be limited to variations in the O^{18}/O^{16} ratio since the amount of O^{17} in water is very small (Nier,1950). As this is the practice in the bulk of literature appearing at present (Epstein, Craig, Dansgaard), only variations in the O^{18}/O^{16} ratio will be considered in the present discussion.

Isotopic fractionation of water in nature may be attributed to a number of processes, both physical and biological. The largest and most significant cause of such variations is probably the repeated condensation and evaporation of sea water. Fractionation in this case is due to the different vapour pressures of H_2O^{16} and H_2O^{18}. For vapour and liquid in equilibrium, the vapour phase is enriched in H_2O^{16} and the liquid phase in the heavier H_2O^{18}.

Data concerning the abundance of O^{18} is commonly expressed in terms of absolute atomic per cent (Dansgaard 1961) whereas variations in the O^{18}/O^{16} ratio is expressed in terms of a δ-value.

of H_2O^{16}. The range of reported δ-values for natural waters is from δ = 0 to δ = -50, δ = 0 being ocean water and δ = -50 snow of the Greenland ice cap (Sharp 1960). The δ value of -50 indicates a number of successive distillations.

The explanation of variations in the natural abundance of O^{18} in ice and snow concerns itself with two processes:

1) a Raleigh distillation process that deals with isotopic composition as a function of the amount of vapour removed or condensed from a given quantity of vapour.

2) variations in the equilibrium constant between liquid and vapour as a function of temperature.

Considering the first effect of distillation, we can state the following equation relating the isotopic composition of the liquid and vapour phase of a distillation system.

$$\underset{\text{Liquid}}{H_2O^{16}} + \underset{\text{Vapour}}{H_2O^{18}} \rightleftharpoons \underset{\text{Liquid}}{H_2O^{18}} + \underset{\text{Vapour}}{H_2O^{16}}$$

The equilibrium constant for this reaction is written

$$K = \frac{[H_2O^{16}\ \text{vapour}]\ [H_2O^{18}\ \text{liquid}]}{[H_2O^{16}\ \text{liquid}]\ [H_2O^{18}\ \text{vapour}]}$$

or
$$K = \frac{\left(\frac{H_2O^{18}}{H_2O^{16}}\right) \text{liquid}}{\left(\frac{H_2O^{18}}{H_2O^{16}}\right) \text{vapour}}$$

This expression for K is alternately referred to in the literature as α = the fractionation factor at the equilibrium temperature. Epstein (1962) has quoted the following equation relating the isotopic composition of water enriched by evaporation to the fraction of the initial volume of liquid remaining:

$$\frac{R}{R_o} = f^{(1-\alpha)}$$

where
$$\alpha = \left(\frac{H_2O^{18}}{H_2O^{16}}\right) \text{liquid} \Big/ \left(\frac{H_2O^{18}}{H_2O^{16}}\right) \text{vapour}$$

$$R = \frac{H_2O^{18}}{H_2O^{16}} \quad \text{of the residual water}$$

$$R_o = \frac{H_2O^{18}}{H_2O^{16}} \quad \text{of the initial water}$$

f = fraction of the initial volume of liquid ramaining. To put this expression in terms of δ-values:

$$\delta = \left[\frac{\left(\frac{H_2O^{18}}{H_2O^{16}} \right)_{\text{sample}}}{\left(\frac{H_2O^{18}}{H_2O^{16}} \right)_{\text{standard}}} - 1 \right] 1000$$

The fractionation equation can be rewritten:

$$\frac{\dfrac{R}{\left(\frac{H_2O^{18}}{H_2O^{16}} \right)_{\text{Std.}}}}{\dfrac{R_o}{\left(\frac{H_2O^{18}}{H_2O^{16}} \right)_{\text{Std.}}}} = f^{(1-\alpha)}$$

Since $\delta = \left[\dfrac{R}{\left(\frac{H_2O^{18}}{H_2O^{16}} \right)_{\text{Std.}}} - 1 \right] 1000$

$$\frac{R}{\left(\frac{H_2O^{18}}{H_2O^{16}} \right)} = \frac{\delta}{1000} + 1$$

and

$$\frac{\frac{\delta \text{ liquid}}{1000} + 1}{\frac{\delta \text{ original}}{1000} + 1} = f^{(1-\alpha)}$$

If the value of the initial liquid is defined as being equal to zero,

$$\frac{\delta \text{ liquid}}{1000} + 1 = f^{(1-\alpha)}$$

or $$\delta \text{ liquid} = [f^{(1-\alpha)} - 1\ ]\ 1000\ .$$

To find the δ-value of the associated vapour phase

$$\alpha = \frac{\left(\frac{H_2O^{18}}{H_2O^{16}}\right) \text{liquid}}{\left(\frac{H_2O^{18}}{H_2O^{16}}\right) \text{vapour}} = \frac{\left(\frac{H_2O^{18}}{H_2O^{16}}\right) \text{liquid} \Big/ \left(\frac{H_2O^{18}}{H_2O^{16}}\right) \text{Std.}}{\left(\frac{H_2O^{18}}{H_2O^{16}}\right) \text{vapour} \Big/ \left(\frac{H_2O^{18}}{H_2O^{16}}\right) \text{Std.}}$$

$$= \frac{\frac{\delta \text{ liquid}}{1000} + 1}{\frac{\delta \text{ vapour}}{1000} + 1}$$

$$\alpha \left(\frac{\delta \text{ vapour}}{1000}\right) + \alpha = \frac{\delta \text{ liquid}}{1000} + 1.$$

Since α is close to 1, the following approximation can be made:

$$\frac{\delta \text{ vapour}}{1000} + \alpha \simeq \frac{\delta \text{ liquid}}{1000} + 1$$

so the δ-value of the vapour is given by:

$$\delta \text{ vapour} = \delta \text{ liquid} + [1-\alpha]\ 1000\ .$$

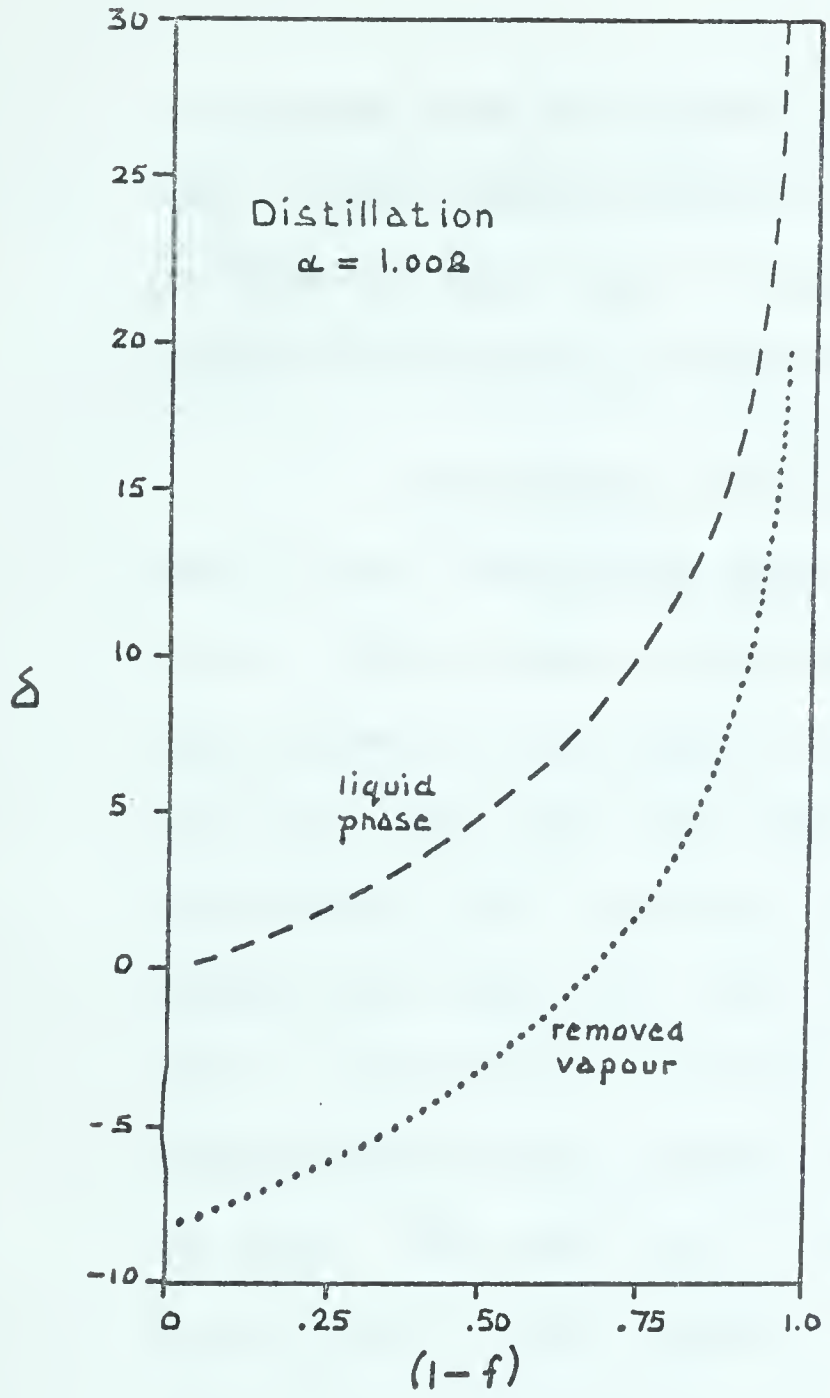
Extending the distillation equation to include condensation from a vapour,

$$\delta \text{ liquid} = [f^{(\alpha-1)} - 1] \ 1000$$

$$\delta \text{ vapour} = \delta \text{ liquid} + [1-\alpha] \ 1000 \quad \text{as before.}$$

These equations in δ produce the plots shown in figure 1.

The α factor quoted here is for water in equilibrium at 25° C. The plots show that the vapour for all fractions is enriched by approximately -8 ‰ in H_2O^{16} over the remaining or condensed liquid for a temperature of 25° C. Considering first the distillation equation, the very first fraction of vapour to be removed is enriched in the lighter isotope to produce a negative δ-value. The δ-value for the vapour gets increasingly positive as evaporation is continued, but in actual fact only the first fraction need be considered as we can assume the ocean to be an "infinite source" of water, and the fraction of water removed stays very close to zero. The δ-values for water from natural sources must always be negative since sea water is the "heaviest" natural water. The equation quoted above assumes that the vapour phase is being removed continually from the system. This is a rather debatable assumption, which is discussed at great length by H. R. Craig (1963). Applying the distillation equation to condensation, the assumption that the liquid phase



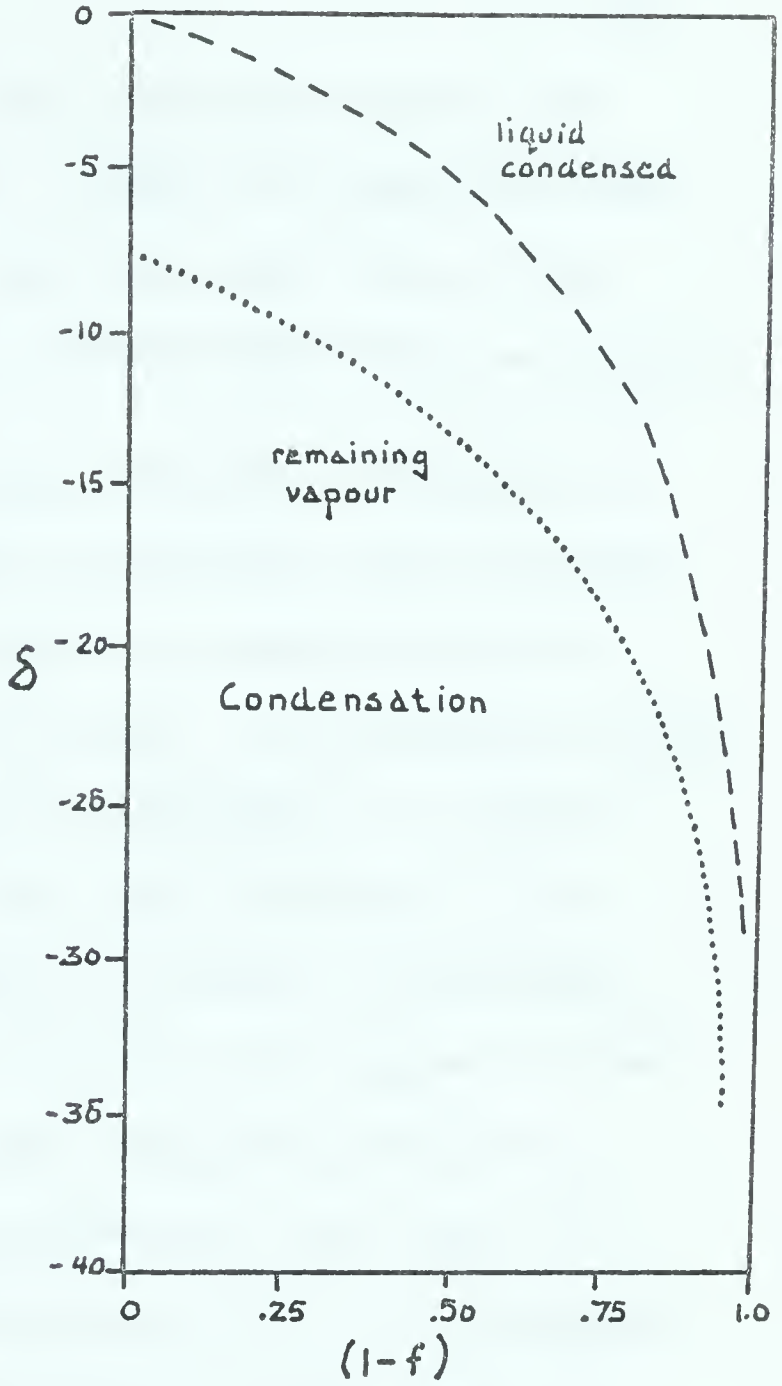


Figure 1.

Distillation and Condensation Curves

is removed from the system is a far more valid one in the case of the precipitation of snow. Again, we have the first fraction of snow with a δ-value very near sea water, this becomes increasingly negative as more and more snow is removed.

Considering now the actual physical situation, a mass of air containing water vapour differing from sea water by δ = -8‰ is blown inland and starts to precipitate the water vapour in the form of rain or snow. Precipitation from this air mass, near the coast at low altitudes in a marine environment will represent the very first fraction to be removed and hence the least negative δ-values. Subsequent motion inland and an increase in altitude will cause further precipitation and progressively more negative δ-values in the snow. The per cent of precipitation in this case is a function of a great number of variables, but is nevertheless positively correlated to the altitude of the air mass. (Dansgaard, 1953). The gradient of δ with altitude depends to a great extent on the physical environment of the snow accumulation area and the direction of the prevailing wind, but it has been conclusively demonstrated by Epstein and co-workers (1960).

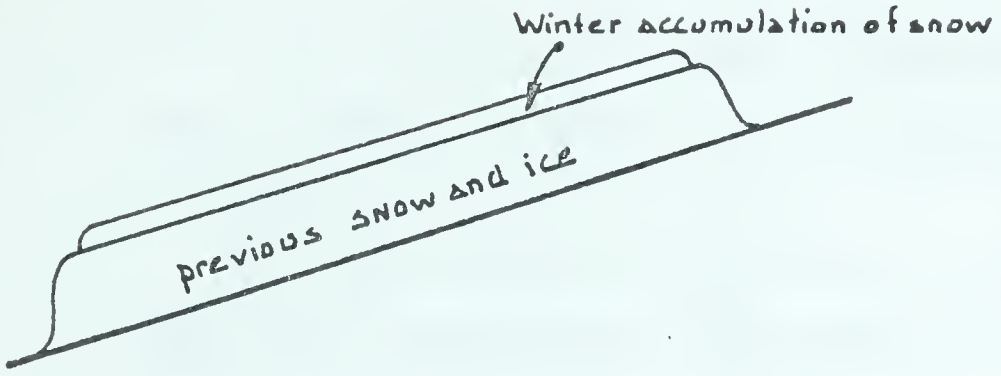
A second effect, the dependence of δ on temperature, has been investigated by a number of workers: Riesenfield and Chang, Wahl and Urey, and Epstein and Friedman. Values

produced by Riesenfield and Chang in 1936 are frequently quoted:

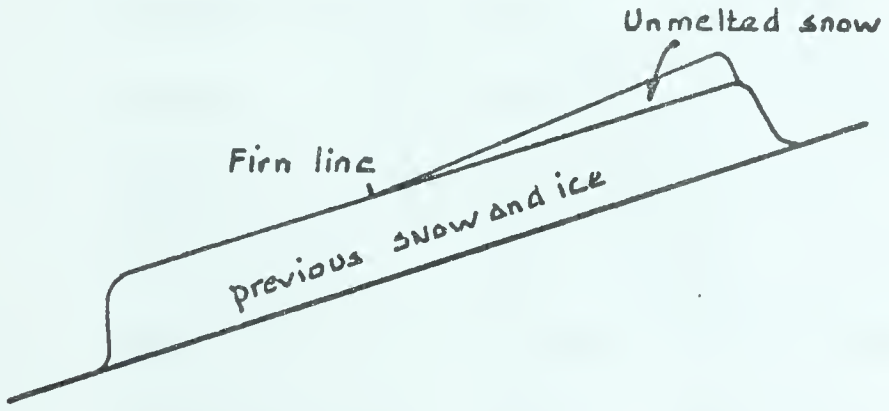
$$\alpha = \frac{\left(\frac{H_2O^{18}}{H_2O^{16}}\right)_{\text{liquid}}}{\left(\frac{H_2O^{18}}{H_2O^{16}}\right)_{\text{vapour}}} = \frac{\alpha \quad 1.005 \quad 1.008 \quad 1.010}{\text{Temp.} \quad 100^{\circ}C \quad 25^{\circ} \quad 0^{\circ}}$$

In other words, throughout the range of climatic temperatures, α changes by about .0003 per degree centigrade. Since present simultaneous collection mass spectrometric techniques can measure α variations of the order of $\pm$ .0001, the temperature effect can be considered to be relatively large. Mechanisms of warming and cooling of an air mass will not be considered, but it will be true that vapour condensed or snow precipitated from a "cold" air mass will be enriched in H_2O^{18} compared to snow from a "warm" air mass.

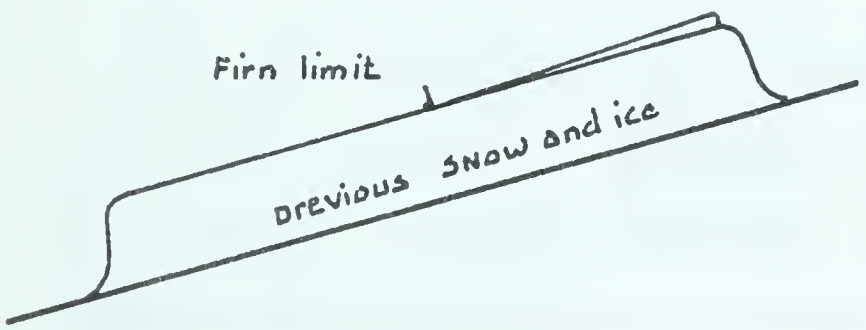
Bearing in mind effects due to altitude and temperature, attention must now be drawn to the morphology of a glacier. In Figure 2 seasonal changes in climatic conditions and snow load are illustrated by a highly idealized block of glacier ice, for a "temperate" glacier.



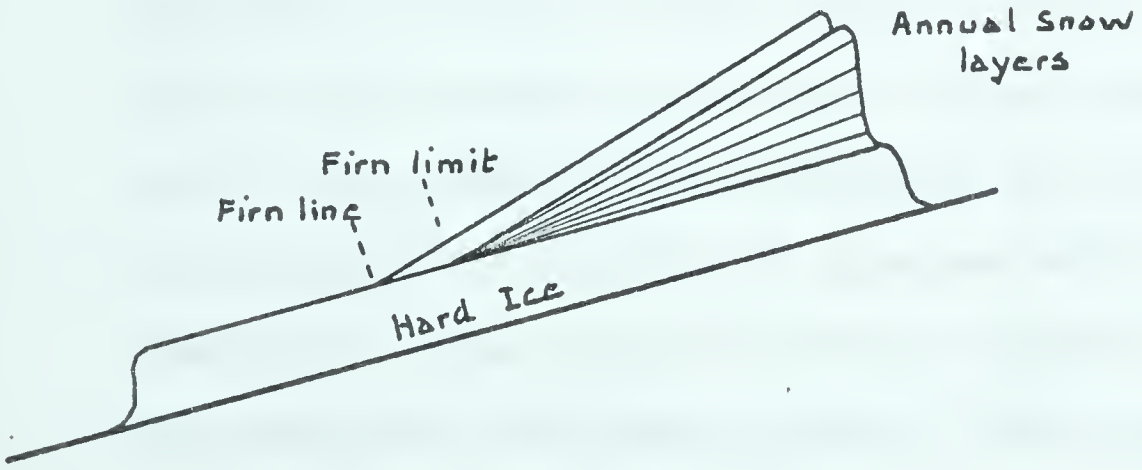
a. Accumulation during the winter prior to melting.



b. Ablation during the spring and summer.



c. Furthest extent of ablation.



d. Overall picture sometime before the furthest extent of ablation.

Figure 2 Idealized diagrams of annual accumulation and ablation of ice and snow on a glacier.

In an idealized valley glacier, we can expect the following conditions to prevail:

1) The terminus will be at a considerably lower altitude than the accumulation area and as a result, there will exist a thermal gradient from the bottom to the top. The combination of thermal gradient and altitude gradient produce variations in the $\frac{O^{18}}{O^{16}}$ ratio known as the altitude effect. The δ-values in the snow get progressively more negative from the firn line to the far reaches of the accumulation area. Such an altitude effect has been measured for a number of glaciers, notably the Blue glacier in Washington where δ varies an average of -.5 per 100 meters. (Epstein 1960).

2) The temperature of deposition will be expected to vary from snowfall to snowfall, as the winter progresses. If we assume no isotopic exchange between layers of snow and between the snow and the atmosphere, δ-values corresponding to temperature variations will be preserved in the snow, providing a record of the winter thermal regime. This record will remain in the snow layers until subsequent melt action has produced homogenization by the downward percolation of melt water. In a "cold" glacier which has an average temperature less than $0^{\circ}C$, annual temperature variations are preserved from year to year. Significant studies on

the Greenland ice cap by Epstein and Benson have produced clearly defined temperature layers which correspond well with stratigraphic studies.

Figure 3 illustrated the variation in δ-values in the precipitated snow as deposition continues throughout the winter.

The δ-values for a number of annual layers observed by Benson (1960) for samples collected from a pit in Greenland are shown in figure 4. It is noted that the coldest temperatures and the lowest δ-values occur during the early spring in each case.

For snow on a temperate valley glacier the temperature scale is partly preserved throughout one summer's melt action, but does not remain past the second year. What is left, however, is an average δ-value which will reflect the altitude effect from station to station.

3) The altitude effect can be used as a natural tracer of glacier flow. If the glacier has a bowl-shaped accumulation area, the snow deposited around the periphery of the accumulation area will have a more negative δ-value than snow deposited in the center due to the altitude difference between these two sites. Snow deposited at the edge of the accumulation area becomes ice at the edge of the main body down

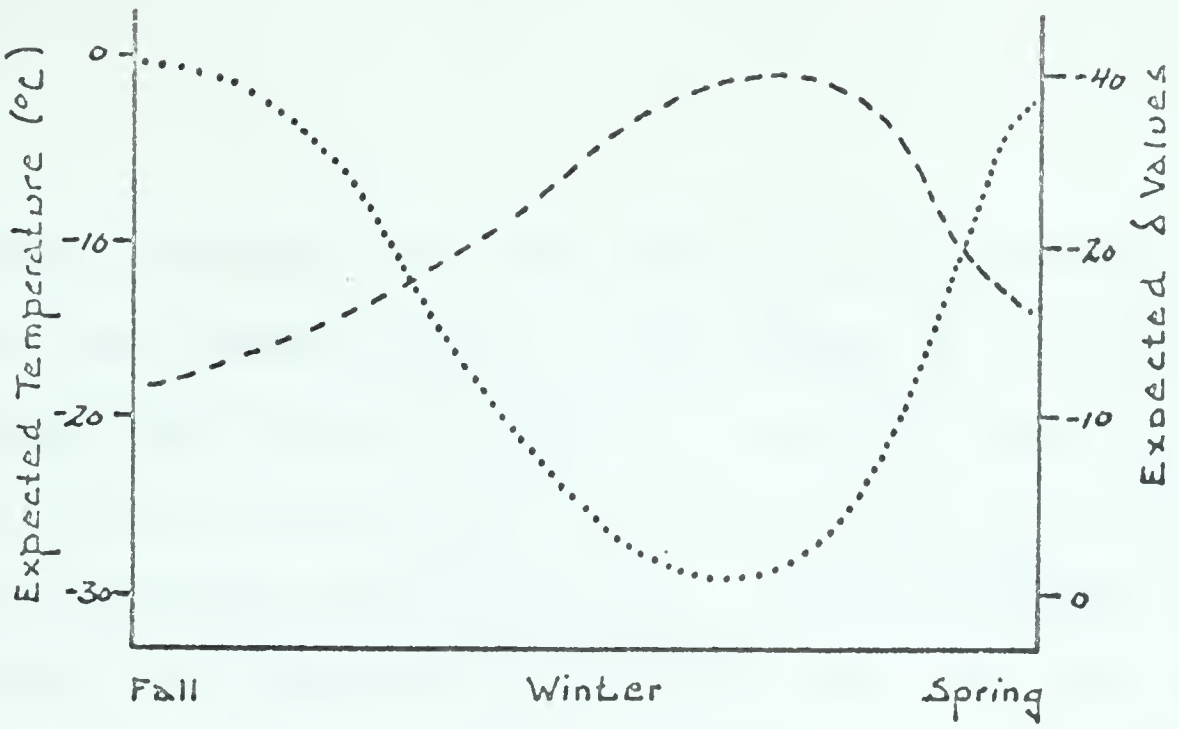


Figure 3 Annual Variations in δ-Values and Temperature as Precipitation Continues During The Winter

..... thermal regime

----- δ-Values

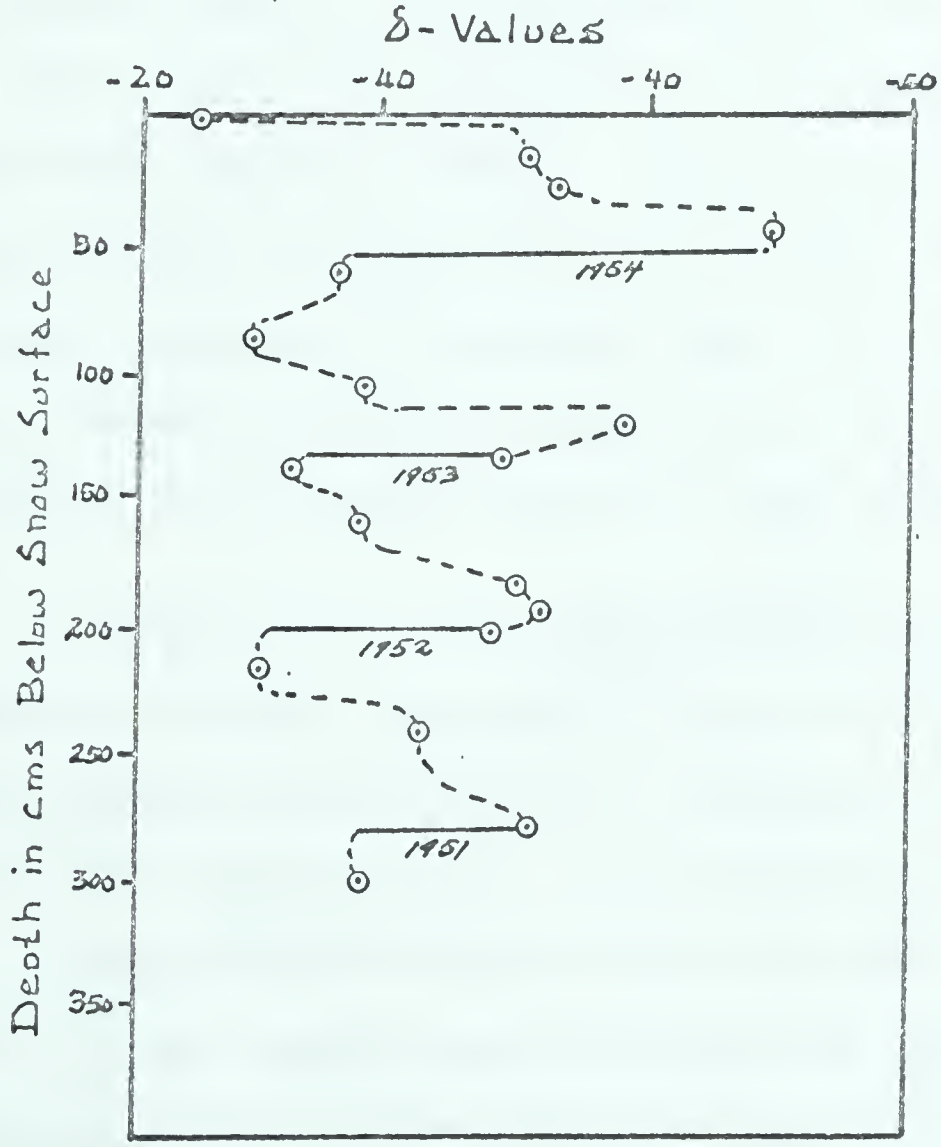


Figure 4 δ-Values For Four Annual Layers Observed in Greenland (Benson, 1960)

glacier, and snow from the center of the accumulation area moves down glacier as ice in the center of the main body. Samples taken below the firn line would be expected to retain the δ-value representative of the altitude from which the snow originally came (figure 5). Hence, δ-values are more negative for the sides of the main body than they are for the center. If a number of tributaries join the main body of ice, transverse variations will be preserved in each stream.

4) The whole mass of ice and its snow load is moving down glacier at a rate dependent on a great number of variables such as amount of accumulation and shape of the valley wall. One of the first workers to investigate modes of glacier flow was Reid (1896). He suggested that snow from the back of the accumulation area travels down glacier, and eventually appears at the surface as ice at the terminus, and snow deposited slightly above the firn limit appears on the surface as ice below the firn limit as shown in Figure 6.

Since snow in the upper reaches is depleted in the heavier isotope compared to snow at the firn line, a reverse gradient of δ-values will appear in the ice down glacier from the firn line, reflecting the altitude effect above. This has been demonstrated to be the case in studies of the Blue and Saskatchewan glaciers by Epstein et al, (1960). More detailed studies of oxygen isotope ratios are being conducted

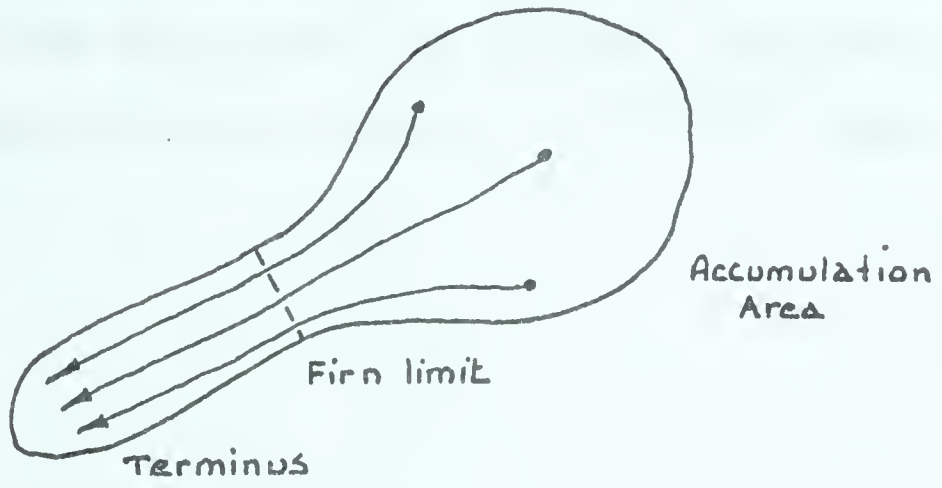


Figure 5

Mode of Glacier Flow in Plan View

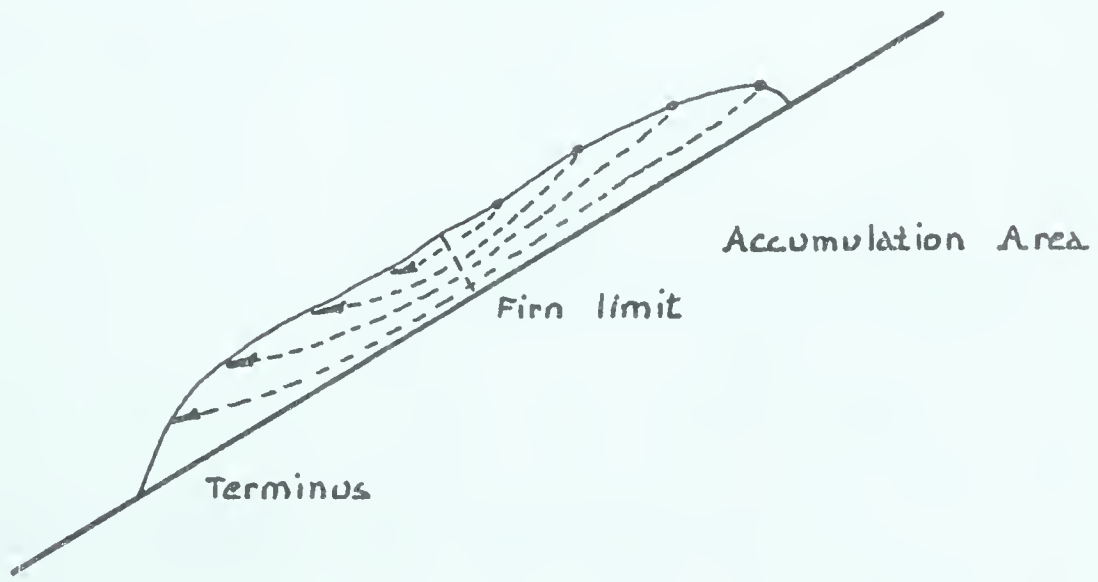


Figure 6

Mode of Glacier Flow in Cross Section

by various workers in an effort to determine the origin of such features as ogives, ice banding, foliation, and different textures of ice by noting the O^{18}/O^{16} ratio.

2. Determination of the O^{18}/O^{16} Ratios

a. Sample Preparation

For various reasons, neither H_2O or O_2 are suitable for direct analysis in the mass spectrometer. It has been the common practice by workers in the field (Epstein, Dansgaard *et al*) to use CO_2 as a working gas. The method consists of equilibrating the water sample with industrial CO_2 at constant temperature and ultimately measuring the ratio of $\left(\frac{CO_2{}^{18}}{CO_2{}^{16}}\right)_{\text{sample}} \Big/ \left(\frac{CO_2{}^{18}}{CO_2{}^{16}}\right)_{\text{standard}}$ by simultaneous collection. We can demonstrate that at equilibrium the δ-value calculated for the CO_2 system will be the same as the required δ-value for the water sample by the following considerations:

$$\delta \text{ for water sample} = \left[\frac{\left(\frac{O^{18}}{O^{16}}\right)_{\text{sample}}}{\left(\frac{O^{18}}{O^{16}}\right)_{\text{standard}}} - 1\right] (1000)$$

$$= \left[\frac{\left(\frac{O^{18}}{O^{16}}\right)_{\text{Sample}} - \left(\frac{O^{18}}{O^{16}}\right)_{\text{Std.}}}{\left(\frac{O^{18}}{O^{16}}\right)_{\text{Std.}}}\right] (1000) \quad \text{i.e.}$$

but it will be true for the equation

$$CO_2{}^{16} + 2\ H_2O^{18} \rightleftharpoons CO_2{}^{18} + 2\ H_2O^{16}$$

that there will exist a fractionation factor α between the two phases such that $\left(\frac{O^{18}}{O^{16}}\right)_{\text{water}} = \alpha \left(\frac{O^{18}}{O^{16}}\right)_{CO_2}$ for a system in equilibrium. Using this in the δ equation for the water system

$$\delta = \left[\frac{\alpha\left(\frac{O^{18}}{O^{16}}\right)_{CO_2} - \alpha\left(\frac{O^{18}}{O^{16}}\right)_{CO_2 \text{ standard}}}{\alpha\left(\frac{O^{18}}{O^{16}}\right)_{CO_2 \text{ standard}}} \right] (1000)$$

The α factor drops out and it will be true that

$$(\delta \text{ water})_{\text{system}} = \left[\frac{\left(\frac{O^{18}}{O^{16}}\right)_{CO_2 \text{ sample}}}{\left(\frac{O^{18}}{O^{16}}\right)_{CO_2 \text{ standard}}} - 1 \right] \times (1000)$$

$$= (\delta \ CO_2)_{\text{system}}.$$

Certain corrections for the initial isotopic composition of the CO_2 and its water content must be made. These will be discussed under methods of sample analysis.

The method of equilibrating the water sample with CO_2 used in this study is outlined below.

1. The bottles shown in the accompanying diagram were sealed to a vacuum system and pumped for a few hours to remove water vapour from the entire system. A fast torching seemed to help this process greatly.

2. The pH of the water sample was adjusted with a small drop of dilute HNO_3 to 6 or lower to facilitate rapid equilibration. (H. C. Urey, 1940).

3. The stopcock on the breakseal was closed, the top of the bottle was removed, the sample rapidly inserted with a dry 25 ml. pipette, and the bottle was sealed up again.

4. The water sample was frozen with a dry ice-alcohol bath, and the bottle was pumped for five minutes to remove the air and CO_2.

5. The sample was melted, brought up to room temperature to drive off trapped gas in the ice, and refrozen.

6. The bottle was pumped for one minute to remove any residual gas driven off by the melting.

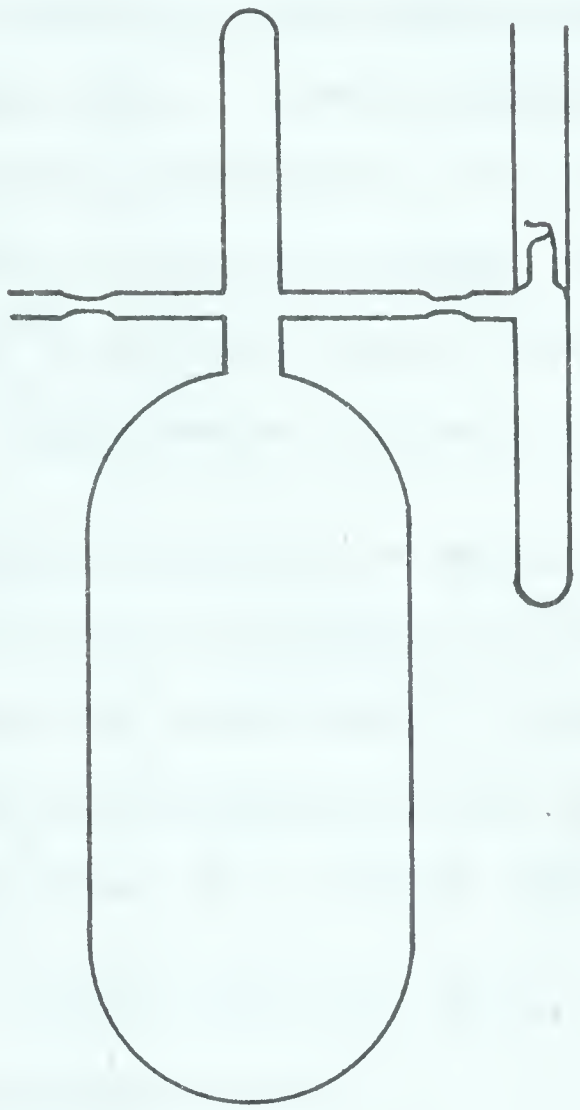
7. Having been thus evacuated, the samples were brought back up to room temperature, and commercial CO_2 was introduced into the bottles to a pressure of about 2 cms. less than atmospheric. A bottle pressure slightly less than atmospheric was found to facilitate easy removal of the bottle from the vacuum system, and easy removal of break seal tubes after equilibration was completed.

8. Equilibration was carried out in a water bath which was thermostatically regulated to 25° C. As much of the bottle was immersed in the water bath as possible to prevent condensation of the water sample in the neck of the bottle and in the stopcock arm (condensed water would be forced into the breakseal upon opening the stopcock to let in the sample). Sample bottles were shaken frequently (as Epstein [1953] points out) "to insure that water deficient in the heavier isotope is not condensed on the walls of the vessel."

9. After equilibration for three days at 25° C a 7 cc. aliquot was let into the breakseal, and the stopcock was closed again. The sample was removed by freezing down the CO_2 in the breakseal and heating the neck to constrict and close it.

A survey of the literature shows that various lengths of equilibration time are used. Dansgaard (1961) finds that two hours is sufficient whereas Epstein (1953) equilibrates samples for three days.

Samples were equilibrated for three days, since the general method of sample preparation outlined by Epstein was used in this study.



Samole Preparation

Bottle

Figure 7

Evacuation of the bottles and sample handling could have been more efficient had glass stopcocks been used rather than glass seals. However, glass seals were used to prevent possible contamination of the sample by water and hydrocarbons in stopcock grease. For similar reasons, metal valves or mercury valves were used for sample handling on the mass spectrometer wherever possible.

Certain corrections must be applied for the amount and content of oxygen in the industrial CO_2 used for equilibration as well as for the water which it might contain. Expressions for these corrections will not be derived here but are quoted from a paper by H. Craig (1957).

Define δ = actual corrected value

δ^1 = measured value

$$\delta = \delta^1 \left(\frac{\rho+\alpha}{\rho}\right) - \frac{\alpha}{\rho} \left(\delta\, O^{18}\right)_{\text{tank } CO_2}$$

where

$$\alpha = \frac{R_{18}\,(CO_2)}{R_{18}(H_2O)} \cong 1.039 \text{ at } 25^\circ \text{ C.}$$

$\delta(O^{18})$ = O^{18} ratio of tank CO_2. Since the tank CO_2 has been used as a reference standard for all analyses, $\delta(O^{18}) = 0$.

ρ = ratio of gram atoms of oxygen in the water sample to the gram atoms in the CO_2. This has an approximate value of = 2.18×10^2 for 25 mls. of sample and 175 mls. of CO_2.

Applying these values of $\rho = 2.18 \times 10^2$, $\alpha = 1.039$ to the equation of δ,

$$\delta = \delta^1 + \frac{\alpha}{\rho}$$

$$= \delta^1 + \frac{1.039}{2.18 \times 10^2}$$

$$= \delta^1 + 4.8 \times 10^{-3}$$

A correction of this order of magnitude is well within the limits of experimental accuracy and will be neglected for the present discussion. A further correction has been suggested by Craig (1957) for the water content of the industrial CO_2 used for equilibration. A scan of the mass units up to mass 46 shown in figure 9 indicates the high percentage of water contamination. With this in mind, CO_2 used for equilibration was first purified by freezing down the water vapour in a dry ice and alcohol bath, and then freezing down the CO_2 with liquid air; the water trap was brought back to room temperature, then the whole system was pumped on to remove any contaminants. After this procedure, it was felt that a correction due to water contamination was unnecessary.

b. Instrumentation

A precision of ± 0.1 of a δ value was obtained on the mass spectrometer by the utilization of a simultaneous collection technique. A 90 degree, 12 inch radius magnetic analyzer was used. Slits were set up for simultaneous collection of ion currents of masses 44 and 46. Ion current measurement and amplification was done by two vibrating reed electrometers as illustrated in figure 8. Measurement of the ratio of the output voltages from the two vibrating reed electrometers was made with a five figure integrating digital voltmeter-ratiometer. The ratios of the two voltages, which were in turn proportional to the 44/46 ion current ratios, were then periodically recorded on a five figure digital printer. The system described proved to be superior to the null method of measurement described by Dansgaard and Epstein, (1962), as it monitors instrument drift very closely and is more convenient. A sketch of the ratio recording system is included below.

Rapid intercomparison of sample and standard was made possible by a magnetic valve switching system described by Thode and Wanless (1953).

The abundance of masses 44, 45 and 46 (Dansgaard, 1961) are as follows:

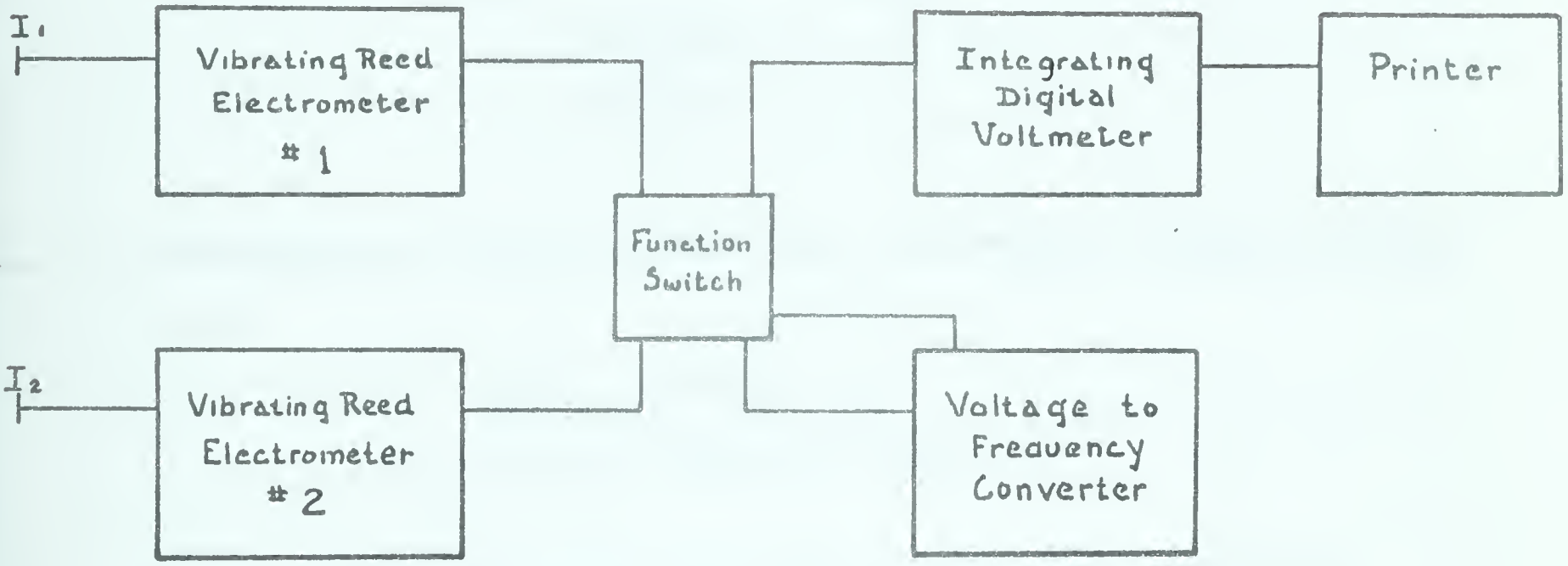


FIGURE 8

BLOCK DIAGRAM OF SIMULTANEOUS ION

COLLECTION MEASUREMENT CIRCUIT

$$44 - C^{12}O^{16}O^{16} = .984$$
$$45 - C^{13}O^{16}O^{16} = .011$$
$$- C^{12}O^{16}O^{16} = 8 \times 10^{-4}$$
$$46 - C^{12}O^{16}O^{18} = 0.004$$
$$- C^{13}O^{16}O^{17} = 9 \times 10^{-6}$$
$$- C^{12}O^{17}O^{17} = 1.5 \times 10^{-7}$$

(presupposing the true fractional occurences of these isotopes to be:

$$C_{12}:C_{13} = 98.90:1.10$$
$$O_{16}:O_{17}:O_{18} = 99.757:0.039:0.204)$$

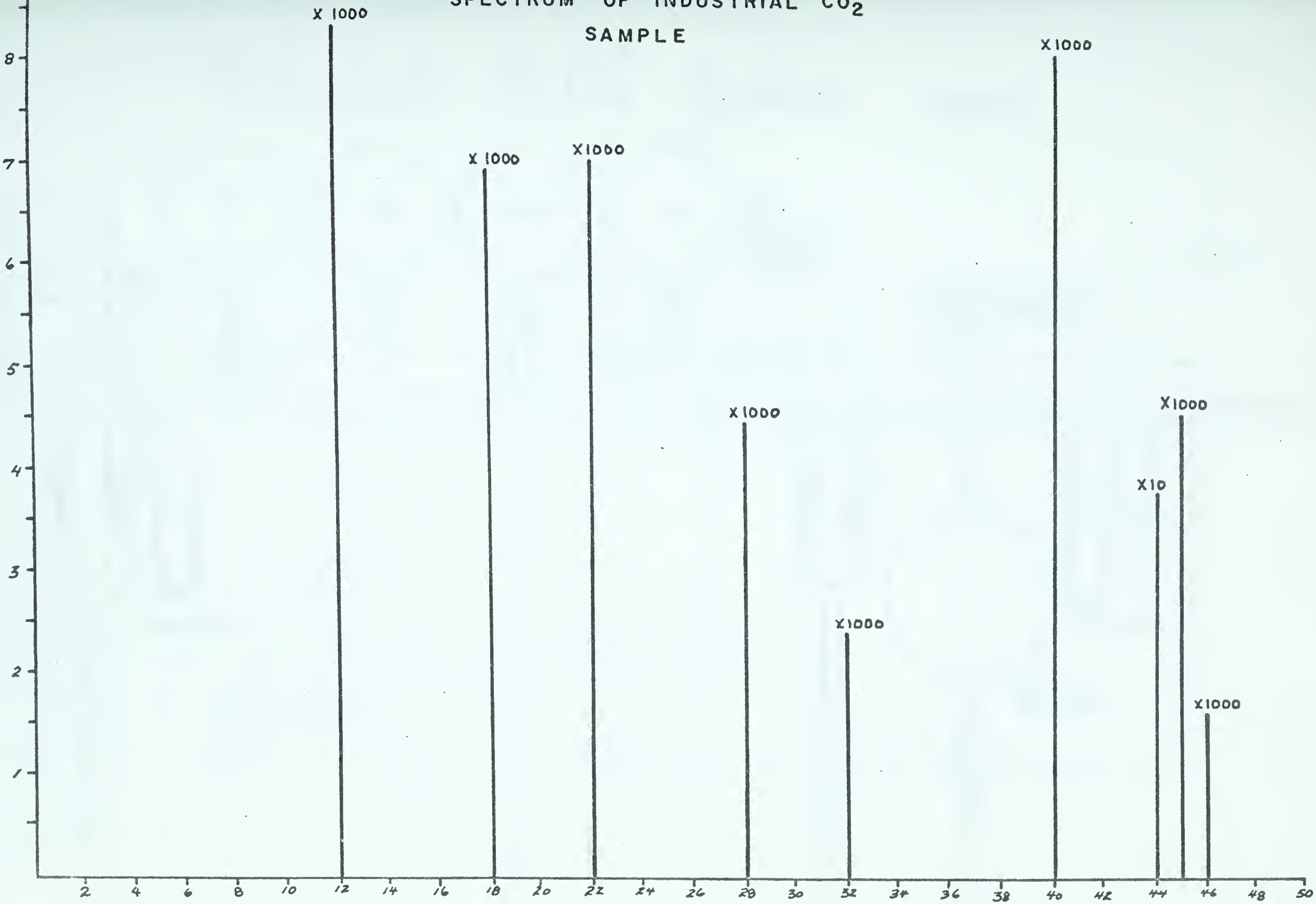
It is evident that only a very small correction must be made for the contributions of $C^{13}O^{16}O^{17}$ and $C^{12}O^{17}O^{17}$ to the mass 46 peak. These are negligible to within the limits of experimental accuracy. A correction due to the presence of C^{13} (which constitutes 1.11% of the total carbon) was avoided by collection of mass 44 and 46 only.

c. Sample Handling System and Sample Analysis

CO_2 samples for analysis on the mass spectrometer were collected in breakseal tubes and installed on the sample manifold (figure 11). After pumping to a suitable pressure to reduce the residual on the mass 44 peak to within 0.40% of the normal sample running pressure, an aliquot of sample and standard was introduced into the respective mercury

SPECTRUM OF INDUSTRIAL CO_2

SAMPLE

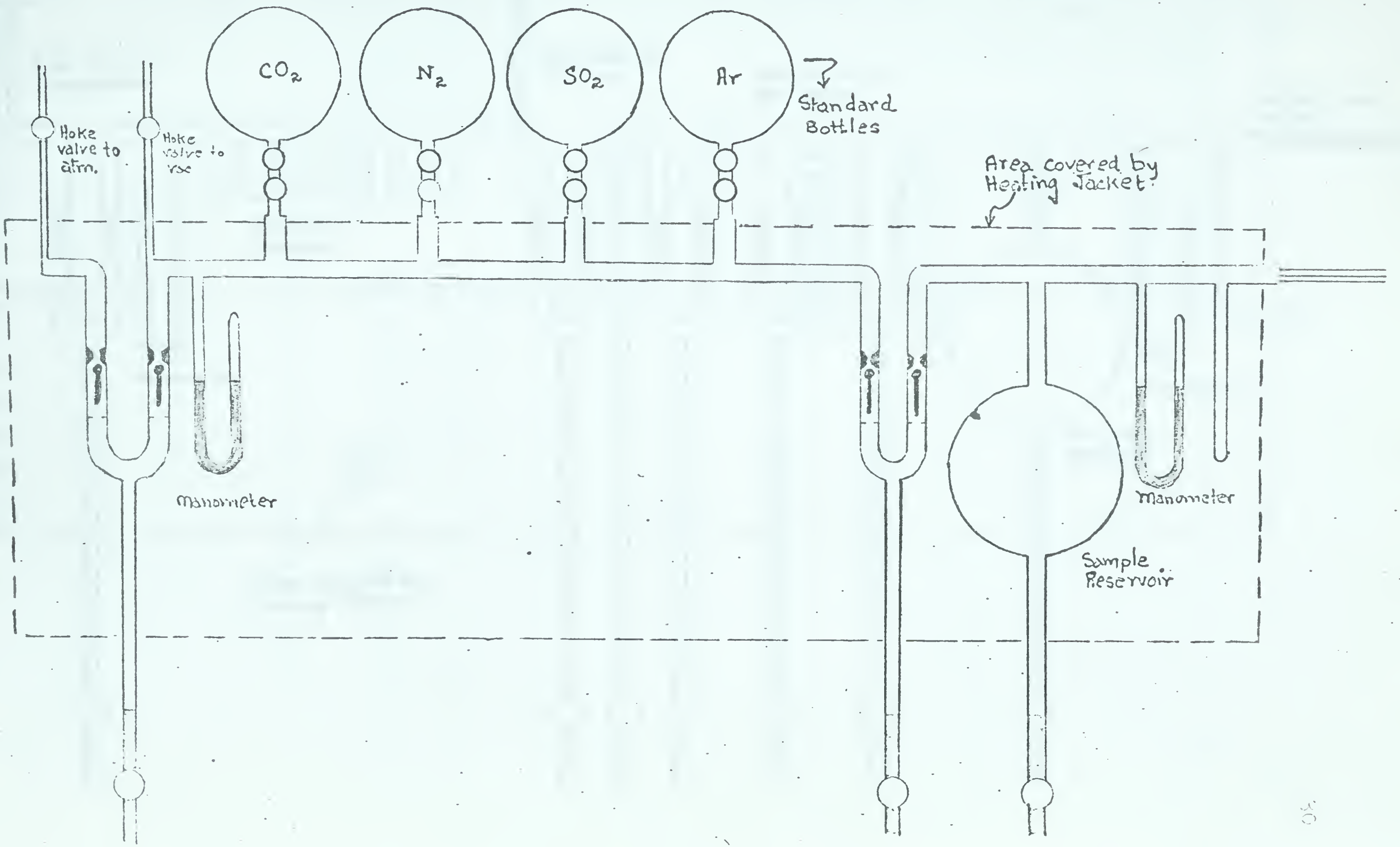


MASS NUMBER

Figure 9

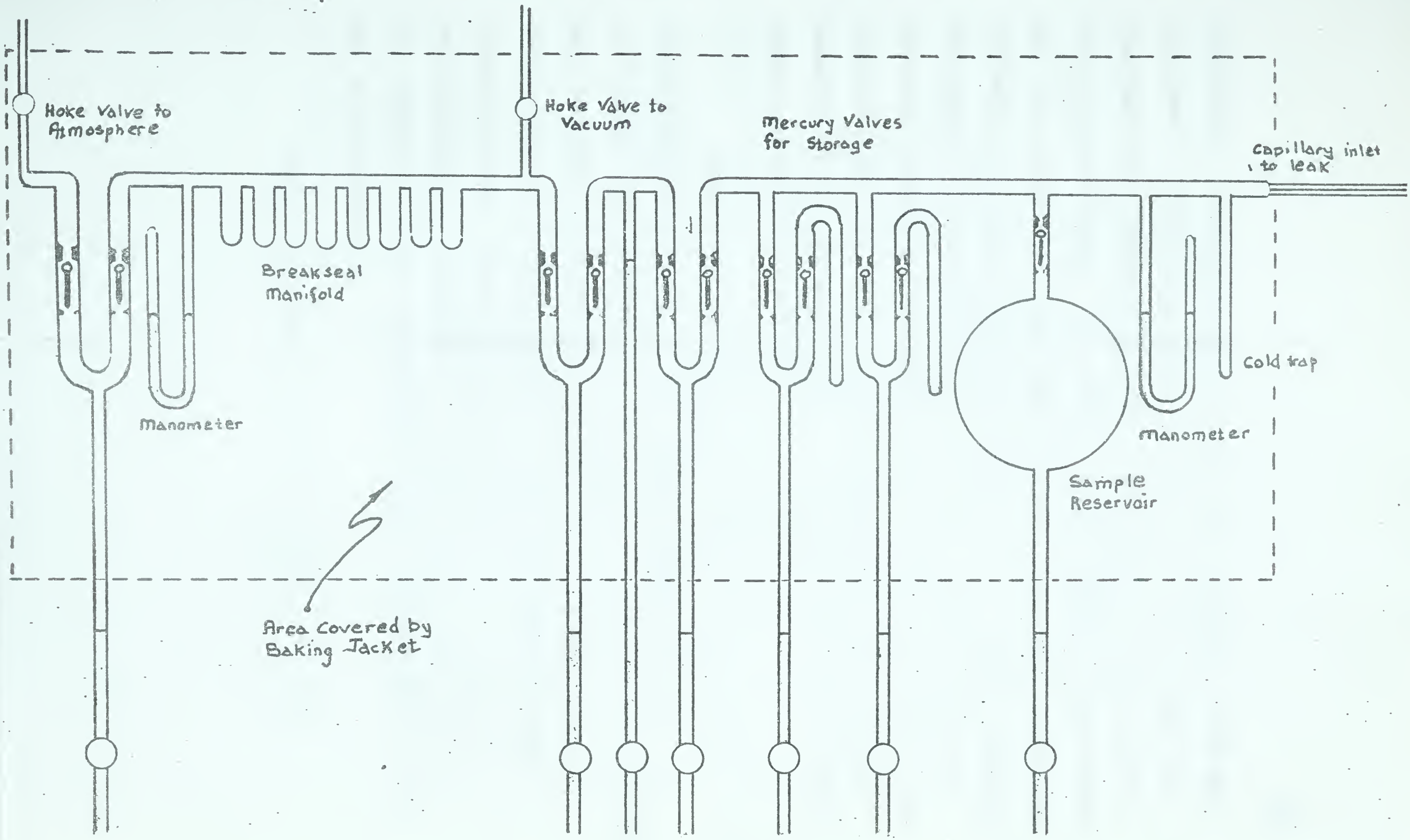
Standard Inlet System

Figure 10



Sample Inlet System

Figure 11



reservoirs. The CO_2 pressure behind the capillary leaks was adjusted by raising the mercury level until both sample and standard line produced the same ion current on single collection of the 44 peak. The exhaust on the magnetic valve system was then adjusted so that switching from one line to the other produced as little as possible change in the "signal strength." Throughout the run the mercury reservoir pressure had to be continually adjusted to maintain the "signal strength" at its initial value and reduce drift in the mass 44/mass 46 ratio measured by the ratio recorder. The sample pressure in each reservoir amounted to about 1 cm. of mercury.

Measurements of the mass 44/mass 46 ratio were conveniently printed by a five figure digital recorder. In each case, approximately ten values of the ratio of the unknown were recorded, then the CO_2 standard was introduced and an additional ten values recorded. This was repeated enough times to insure reproducibility of results. A sample of recorded ratios is given in the table below. Each ratio in the table represents the average of ten recorded ratios from the digital printer.

Sample B-3 run at 10:10 P.M. March 18, 1965:

Standard (1)	0.83651
Sample (1)	0.83269
Standard (1)	0.83673
Sample (2)	0.83268
Standard (3)	0.83645
Sample (3)	0.83237
Standard (4)	0.83609

Since a certain time drift was present in each run, a given sample ratio was compared with the time average of the standard ratios run immediately before and after than sample. As an example, the results for sample B-3 were arrayed and calculated as follows:

Standard Ratio		Sample Ratio	Rx - Rs	$\frac{Rx - Rs}{Rs}$ = δ-value
(1) .83651				
.83662	(1)	.83269	-.00393	-393/83362 = - 4.70
(2) .83673		.83269	-.00404	-404/83673 = - 4.84
.83659	(2)	.83268	-.00391	-391/83659 = - 4.67
(3) .83645		.83253	-.00392	-392/83645 = - 4.69
.83627	(3)	.83237	-.00390	-390/83627 = - 4.66
(4) .83609				

average δ-value = - 4.71

The value quoted here is calculated as a δ-value relative to the working standard (purified industrial CO_2). The contamination of a sample by previous samples on the manifold plus residual mass 44 peak in the instrument was found to be within the limits of experimental accuracy and was therefore neglected in the calculations of δ. Samples were corrected for drift in the standard reservoir of CO_2 by occasionally analyzing the same sample (stored for this purpose in a mercury cutoff, on the sample line). (Figure 11).

δ-values listed under experimental results have been calculated relative to "Standard Mean Ocean Water" (Craig, 1961) and have an accuracy of ± 0.1 δ. In his paper of 1961, Craig states that the values of NBS-1 standard water sample relative to SMOW is δ = - 7.94. The average value of five determinations of NBS-1 was found to be δ = + 10.76 relative to the secondary standard CO_2 used for analysis. Thus the secondary standard CO_2 had a δ-value of -18.70 relative to SMOW.

Preliminary Studies

Prior to processing glacier samples, certain preliminary investigations were carried out to determine the suitability of the method of sample preparation and the internal consistency of results. δ-values quoted here are relative to "Standard Mean Ocean Water" (SMOW).

1. Reproducibility of Results.

Two identical CO_2 samples were prepared in such a fashion that fractionation would not occur. This formed a test on the method of removal of breakseals from the sample bottles.

δ-values of these CO_2 samples were:

1 - 20.4

2 - 20.4

Six samples were produced to test the method of sample preparation by equilibrating 25 mls of distilled water in each of six equilibration flasks, and removing the breakseals in the usual fashion.

δ-values from this experiment were:

- 19.4
- 19.6
- 19.6
- 19.6
- 19.5
- 19.4 .

This set of values has a standard deviation of $\sigma = \pm 0.1$. It is from this experiment that a precision of ± 0.1 of a δ-value for this present study was obtained.

2. Distillation Experiment.

As was mentioned in the introduction to oxygen isotope variations, one of the methods of producing isotopic fractionation of water is by Rayleigh distillation. To try to produce fractionation in this fashion a column of distilled water was left to evaporate in a room in which there was little circulation of air. The column was 15 cms. diameter by 100 cms. long and was filled with distilled tap water. Evaporation of all but 75 mls. of the column of water took 11 months. Values displayed on the graph (figure 12) have been adjusted to make the δ-value of the initial water equal to zero.

For distillation where the vapour phase is being continually removed, the following Rayleigh distillation equation can be written in terms of the fraction of the original water remaining:

$$\delta \text{ liquid} = [f^{1-\alpha)} - 1]\ 1000'$$

A derivation of this equation is given on page 7 . Since evaporation was carried out at room temperature, a value of $\alpha = 1.008$ was used in the calculation of the theoretical curve.

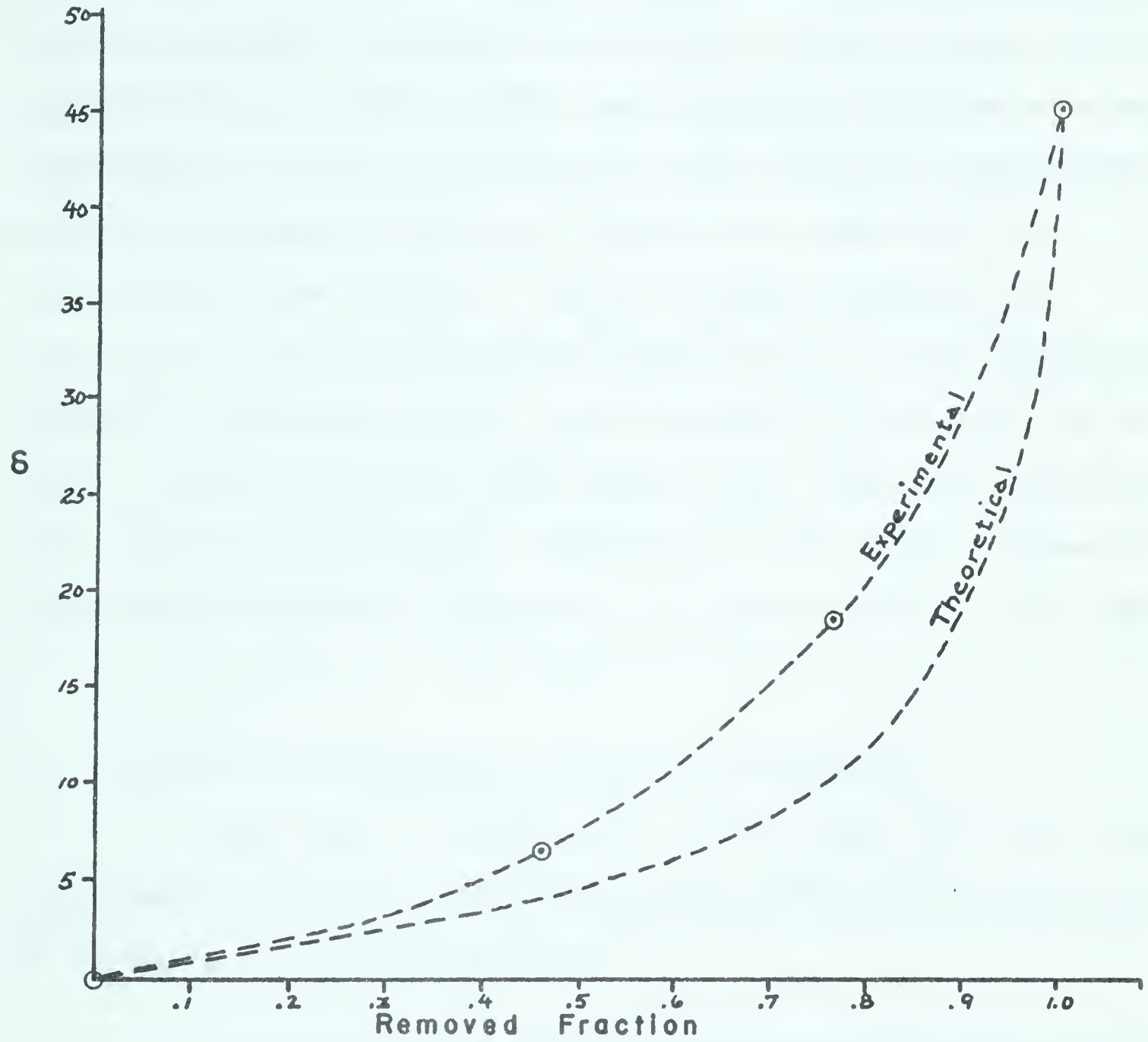


Figure 12

Distillation Curves

Figure 12 shows that a satisfactory agreement was obtained between the theoretical curve and the values produced experimentally. Craig (1963) has performed a series of similar experiments in which he points out that there is considerable isotopic exchange between the evaporating water and the surrounding water vapour. The discrepancy between the theoretical curve for Rayleigh distillation and the experimental values in his experiments is attributed to this source of error. Craig, points out in the same paper that a Rayleigh distillation curve is more closely reproduced if the vapour phase is continually removed by sweeping the system with an inert gas.

4. A General Description of the Glacier System.

Snow and ice samples for this study were taken during the summer of 1963 from the Hubbard and Kaskawulsh glaciers in Canada's St. Elias Mountains.

The Kaskawulsh glacier lies near the Alaska-Yukon border. It flows out toward the interior plateau and the Alaska highway from a snow catchment basin located to the north of Mount Queen Mary. The Kaskawulsh glacier forms the source of the Slims River which flows into Kluane Lake. The adjoining Hubbard system flows from the same general area south into Yakutat Bay and the Gulf of Alaska. Two of the main tributaries of the Hubbard and Kaskawulsh glaciers

have a common accumulation area at an altitude of between 7,000 and 9,000 feet, located at 139°4' longitude and 60°45' north latitude (Figure 13). It was in this accumulation area that part of glaciological investigations were carried out.

The field camp for the summer of 1963 has been described by Ragle (1964). It was situated a few miles on the Kaskawulsh side of the geographical divide separating the Hubbard and Kaskawulsh systems and was supported by a base camp at Kluane Lake.

The general area mentioned above is characterized by a combination of a cold marine climate on the slopes bordering the Gulf of Alaska, and a dry cold climate that dominates the interior plateau. The extent to which the moist Pacific air penetrates the St. Elias mountains is not known, but the extremely large annual accumulation of snow suggests that the Hubbard and Kaskawulsh glaciers are products of deposition of H_2O derived from the Pacific Ocean.

J. M. Havens and D. E. Saarda have published a summary of meteorological data for the summer of 1963 which points to the possibility of the intrusion throughout the accumulation season of the cold dry air from the interior plateau. The predominate wind direction was nevertheless from the south and west throughout the period of observation.

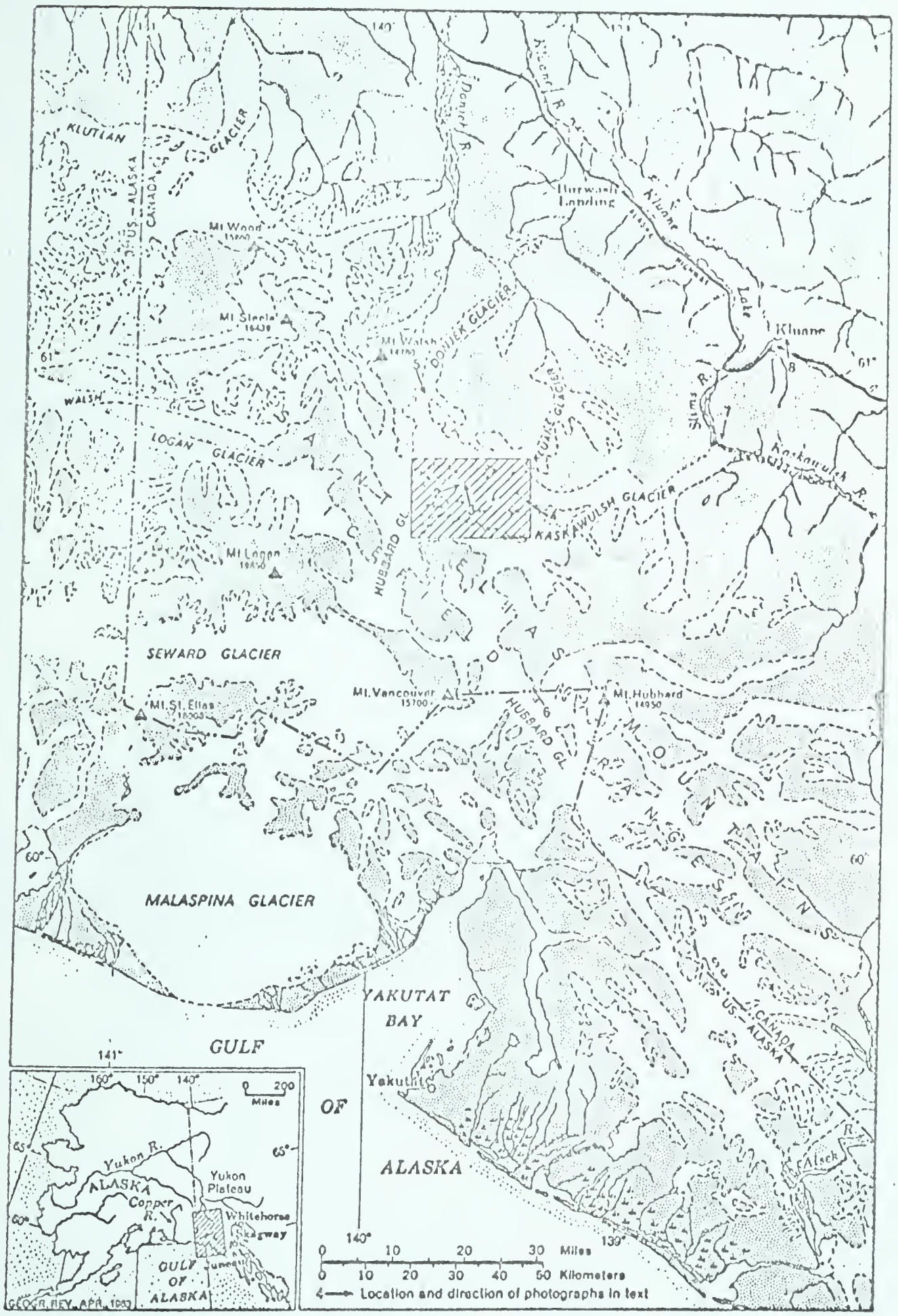


Figure 13 Map of the General Area of Investigation

The shaded area shows the region reproduced in Figure 14.

MAP OF A PORTION OF

THE ICEFIELD RANGES, ST. ELIAS MOUNTAINS

YUKON TERRITORY, CANADA

SCALE 1:30,000

139°30'

139°40'

139°30'

60°45'

60°

K A S K A W U L S H G L A C I E R

(North Arm)

(Central Arm)

(Arm of)

H U B B A R D G L A C I E R

RESEARCH STATION ● 2612 ▲ 2909 2583

60°45'09" N 139°34'14" W

Ground control by Walter A. Wood and Robert W. Mason, 1961
Aerial photography by Royal Canadian Air Force, July, 1951

This map is a product of the Icefield Ranges Research Project under the direction of Walter A. Wood, and sponsored by the Arctic Institute of North America and the American Geographical Society. It is based on stereophotogrammetrical plotting at The Ohio State University under the direction of Arthur J. Brandenberger, and prepared for publication by the American Geographical Society.

ALASKA CANADA
LOCATION OF MAP
Mt. St. Elias
Kluane L.
Yakutat
PACIFIC OCEAN

LEGEND

- Ten-meter contours
- Fifty-meter contours
- Crevasses
- 2574 Elevation (in meters), 1951
- 2608 Elevation (in meters), 1961
- 2643 Elevation (in meters), 1962
- 459 Ice Thickness, 1962
- ▲ Ground-intersected points
- ◎ Photogrammetrically determined points
- o Glaciological stations, 1962
- O Glaciological stations, 1961
  Contours depict the configuration of the glacier topography as it prevailed in 1951

Kilometers 0 1 2 3 4

Miles 0 ½ 1 1½ 2 2½

Copyright, 1963, by The American Geographical Society of New York

Litho by E. Nesbitt & Co., Baltimore, Md.

Figure 14 Map of Accumulation Area

Figure 15 Map of the Kaskawulsh and Hubbard Systems

Glaciological investigations carried out during the summers of 1962 and 1963 have been reviewed by P. Wagner (1963) and W. A. Wood (1963). Wagner and Wood describe the Kaskawulsh as being a temperate valley glacier existing at the pressure melting point and therefore isothermal throughout the budget year. The glacier is approximately 50 miles long from its geographical divide with the Hubbard glacier and varies in width up to 3 1/2 miles depending on the number of tributaries entering the main stream. The Kaskawulsh glacier is fed from three major tributaries which are described as the north, central, and south arms. (Figure 15). The north and central arms extend from the snow catchment basin mentioned above, and the south arm extends from a separate accumulation area somewhat to the southeast of the snow field on which the camp was located. No major ice falls impede the flow of the glacier. The lack of crevasses on the central arm made it the logical choice for a longitudinal sample profile.

Elevations of the various features quoted here are taken from the St. Elias map sheet of the Yukon Territory published by the Department of Mines and Technical Surveys, of the Dominion Government (1961). Part of this map is reproduced in Figure 15.

Elevations:

Terminus	2750'
Firn limit	6500'
Upper camp	8500'
Meteorological station	8650'

Seismological evidence indicates an average thickness in excess of 2,500', and flow data suggests that the north arm is advancing about 300' per year.

Sampling for this project began as soon as logistics would allow the establishment of a camp on the glacier. The first samples were taken shortly after the melt season had begun. The onset of the melt season was indicated by the presence of a few ice layers in the first glaciological pit dug on June 20, 1963. Sampling was continued until August 28, 1963, when colder weather suggested that the height of the melt season was over. Pit studies and the down glacier sample profiles were made with the following ideas in mind:

Samples Taken	Date and Sample Number	Project
22	A-1 to A-22 June 20	First pit study before extensive melt action had taken place. This was continued down to the previous summer layer in an effort to correlate oxygen isotopes with meteorological and glaciological data.

Samples Taken	Date and Sample Number	Project
13	L-1 to L-13 August 29	Auger hole as deep as possible to show annual variations preserved in the yearly layers. Samples were taken from the center of a core brought up with the auger.
3	E-1 to E-3 July 7	Summer snowfall samples correlated with temperature information.
4	B-1 to B-4 June 23	Ice layer study in the first pit after the melt season had progressed to detect downward percolation of surface melt water.
6	C-1 to C-6 June 29	Altitude effect on a nearby slope representing an altitude gradient of over 1000' covering a single snowfall on this slope.
6	F-1 to F-6 July 28	Altitude effect in pits at different altitudes. This should show up in all of the projects mentioned in this list.
3	E-1 to E-3 July 7	Various controls to show the internal consistency of results, also to show any loss of samples due to evaporation.
13	L-1 to L-3 August 29	Re-do a pit near the site of the first pit to show melt action and diagenesis. This project was coupled with the auger hole mentioned in project number 2.

Samples Taken	Date and Sample Number	Project
10	H-1 to H-10 August 22-26	Longitudinal profile down glacier from firn line to the terminus in an effort to correlate surface ice samples with flow theory and try to use oxygen isotopes as a tracer of glacier flow.
9	J-1 to J-9 August 24	Transverse profile across the main body of ice including all three arms to show a possible altitude effect remaining in the ice and to suggest a mode of glacier flow.
16	D-1 to D-16 July 17	Pit study down the Hubbard glacier with the same purpose as project number 2. There might be a possible difference between these two pits if the winter accumulation of snow had a different origin.
1	I-1 August 23	Sample from an ice cored moraine located between the north and central arms.
3	M-1 and M-2 August 14	Water samples from melt streams at the terminus.

Total number of samples = 100.

5. Sampling Methods.

To avoid melting of snow samples taken for density measurements glaciological pits were dug at night. In each case the pits extended at least to the previous summer's surface as evidenced by the stratigraphy and density of the various snow layers. A sketch of the approximate dimensions of a pit is given in figure 16.

While digging the pit, snow was thrown to either side leaving a section of undisturbed snow surface in front. A board was placed on this surface at the edge of the pit to supply a datum for depth measurements. As shown in Figure 16, a working wall was cut in the snow to supply a clean fresh surface fron which samples could be taken.

The method used for collecting samples for O^{18} - analysis involved taking a core of snow from the side of the pit with a 1"-diameter, 20"-long aluminum tube. This tube was open to the air at one end only. The snow core was completely melted by gently heating the tube with a propane torch, taking great care that the snow was just melted and as little water vapour escaped as possible. The water was poured into a 45 ml. plastic bottle and tightly capped. The artificial melting of the snow was necessitated by the size of the bottle. It became readily apparent that a wiser choice of bottle would have been one which would contain enough snow

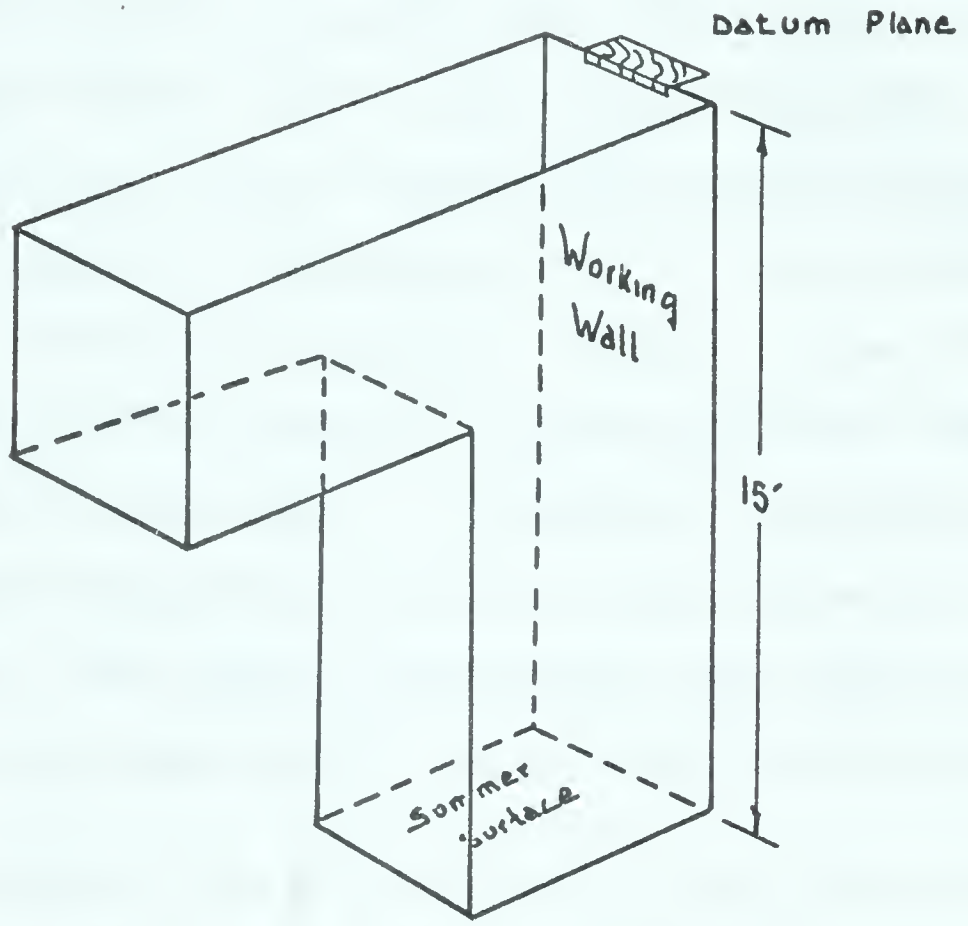


Figure 16

A Sketch of a Glaciological Pit

to give the required 25 mls. of water without heating it first. A wide mouthed bottle proved to be extremely convenient for samples taken by packing the bottle full of snow. The bottles were filled to the top wherever possible to prevent evaporation, and the sides were indented slightly before being capped to demonstrate that if the cap were airtight the indentation remained. Using the idea of indenting the side of the bottle, it was found that very few of the bottles were airtight. To check the evaporation rate, the water level was marked on some of the bottles at the time of filling. In this fashion evaporation was found to be very slight or not observable over a period of one year.

Sampling of hard ice below the firn line presented other problems. Due to the heavy accumulation during the winter, snow near the 1963 firn limit was completely saturated with water, making sampling impossible. As a result, collection of samples of hard ice was started 2 miles down glacier from the firn limit as shown in figure 15. The Kaskawulsh glacier for the most part is not heavily crevassed. This simplified down glacier travel, but melt water streams made it necessary to choose slightly elevated areas as sample sites. The technique used here was to chop away the dirty surface ice making a basin one foot in diameter and a few inches deep. Being thus satisfied that the ice in the basin was clean and relatively dry, the aluminum tube used for snow

coring was filled with ice chips from the bottom of the basin and melted in the usual fashion. Epstein et al. (1960) has found systematic variations in the O^{18} content of the different types of ice. With this idea in mind, only coarse bubbly ice was sampled to avoid introducing more variables into the interpretation of results.

Summer snowfall samples taken on July 7, 1963 were collected by fashioning an aluminum foil funnel and mounting it on a support so that water from the wet snow would drip into the bottle upon melting. No evaporation was expected due to the slight period of time taken to collect each sample.

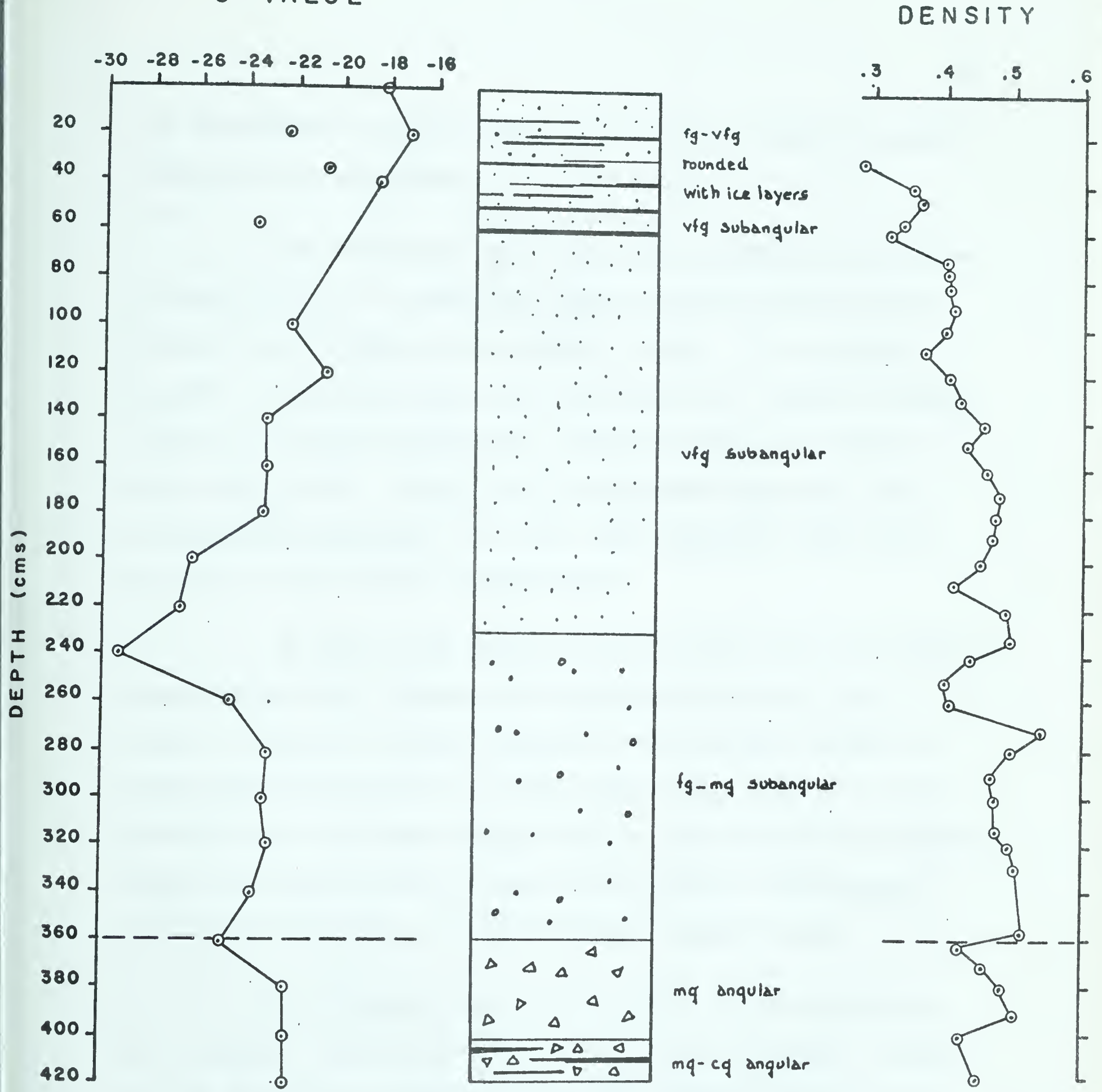
6. Experimental Results and their Proposed Significance

a. Pit Studies on the Hubbard and Kaskawulsh Glaciers

Epstein and Sharp (1960) have described pit studies that show annual trends in a winter accumulation of snow on a number of different glaciers. Due to large accumulation and great size, the Hubbard and Kaskawulsh glaciers were particularly suited for this type of study. Pit number 1 was dug on June 20 on the Kaskawulsh glacier at an altitude of 8396 feet, with the idea of taking samples before the melt season had caused considerable downward percolation of melt water from surface snow. The slight extent of melting up to June 20, 1963 was shown by the lack of ice layers in the pit. The location of the pit is shown on the map reproduced in figure 16.

δ-values from pit number 1 show a marked annual variation. The most negative values occur in snow deposited during the middle of the winter, and the least negative values occur in snow deposited in the spring and fall. The band illustrated in this pit coincides with the idea of the most negative δ-values occurring in snow deposited at the coldest temperatures during the middle of the winter.

Figure 17 illustrates a correlation between oxygen isotope variations and density variations from this pit. From such a correlation it seems possible to postulate that


δ - VALUE
-30 -28 -26 -24 -22 -20 -18 -16
DEPTH (cms)
20 40 60 80 100 120 140 160 180 200 220 240 260 280 300 320 340 360 380 400 420
fg-vfg rounded with ice layers
vfg subangular
vfg subangular
fg-mg subangular
mg angular
mg-cg angular
DENSITY
.3 .4 .5 .6

vfg — less than .5mm

fg — less than 1mm

mg — less than 2mm

cg — greater than 2mm

FIGURE 17

PIT NUMBER 1

Showing the Relationship Between Oxygen Isotopes, Stratigraphy, and Density

the variations in snow density may be due to similar annual trends in the temperature of deposition.

For snow above and below the preceding fall-winter boundary there is a definite oxygen-isotope discontinuity which is also shown in the density values. The average δ-value of samples above the previous years' summer surface (-23.1) is significantly more negative than the δ-values below this layer. This would be expected since the snow for the previous year is at its least negative value near the top of that year's accumulation.

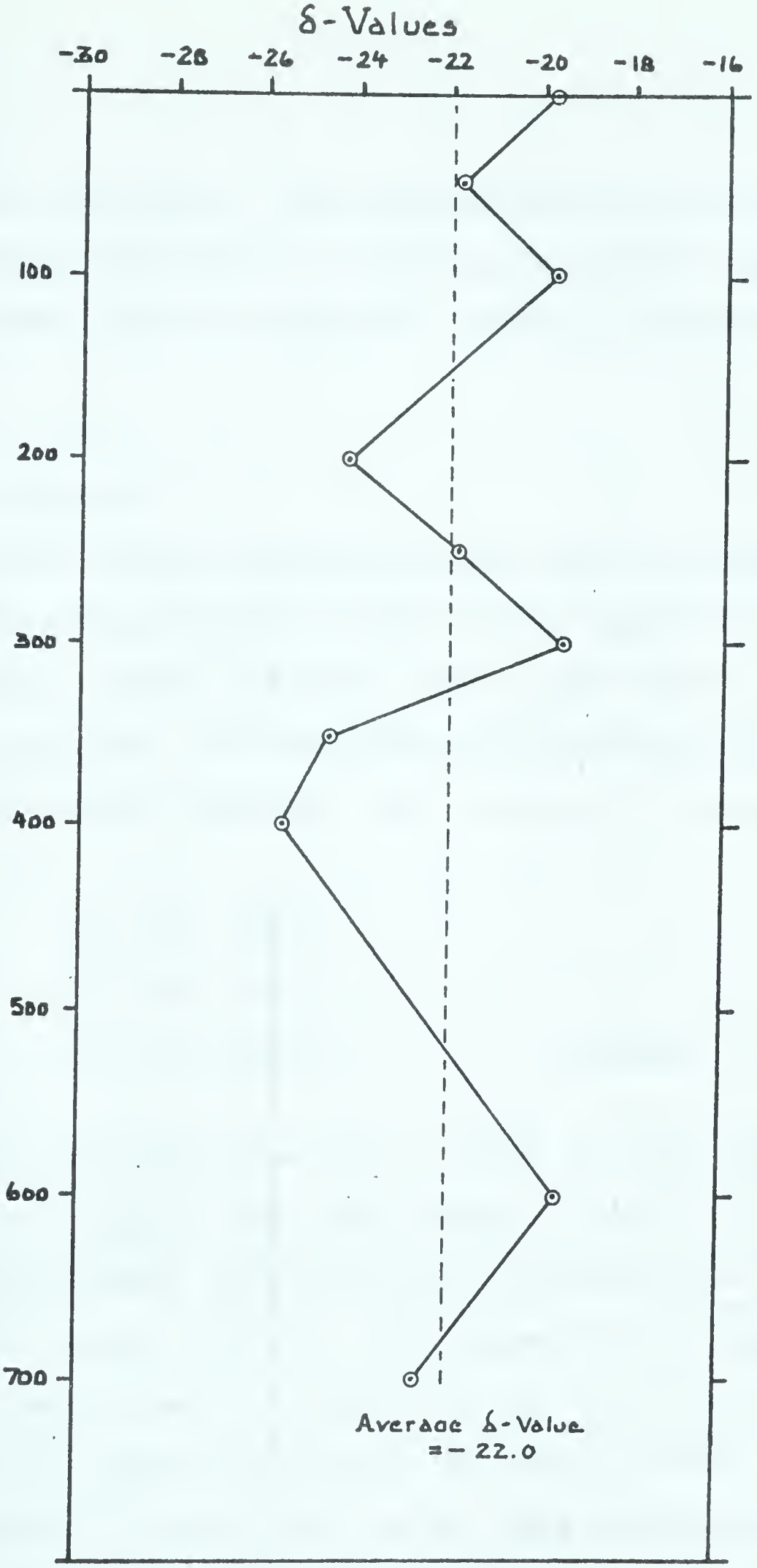
It would seem that the spring snow from the previous year had not been completely homogenized with the snow below it by melt action, and part of the annual trend for that year had remained. In the case of pit number 1 it is difficult to correlate oxygen isotope variations with stratigraphy as the majority of snow shows little stratigraphic variation except for a slight change in grain size.

The average value of $\delta = -23.1$ is quite close to the average δ-value found for ice samples by Epstein (1959) on the Malaspina glacier (in the same general area as the Kaskawulsh Glacier).

A number of ice layers were sampled from pit number 1. Their values are shown along with the δ-values for snow samples in figure 17. Ice layers near the surface of this pit were found to have a more negative δ-value than the surrounding snow. This represents an enrichment in the ice of the lighter isotope. Values from ice layers sampled by Epstein (1959) are scattered both above and below the surrounding snow in such a fashion that it is difficult to say that the formation of ice layers represents a significant fractionation phenomenon.

b. A Borehole at the Site of Pit Number 1.

The site of pit number 1 was re-visited on August 29, after glaciological evidence suggested that the peak of the melt season had passed. In the vicinity of the first pit, a seven meter auger hole was dug with the idea of noting any variation of the δ-values found at the first of the season. δ-values for this study are shown in figure 18. The average δ-value was found to be $\delta = -22.0$. The difference between this value and $\delta = -23.1$ obtained from the first pit study is small enough to be attributed to sampling error. Variations of δ in this auger hole show that the least negative δ-values and the most negative δ-values for one annual accumulation have been removed while the average has remained



Auger Hole at the Site of Pit Number 1.

Figure 18

close to the same value. This phenomenon can be attributed to the downward percolation of surface melt water tending to homogenize and dilute the isotopic ratios in the snow.

c. Summer Rainfall.

Three summer rainfall samples were collected on July 7. The prevailing wind direction, as noted by the meteorological station, (Havens, 1964), was from the east at 2 miles per hour. The temperature of deposition was a few degrees above freezing. The δ-values for these samples were:

1) $\delta = -22.6$

2) $\delta = -22.6$

3) $\delta = -22.5$ average $\delta = -22.6$.

The average value of $\delta = -22.6$ for this study is slightly less negative than the average for the pit studies but is, nevertheless, quite a bit more negative than was expected for summer rainfall. The explanation of this may lie in the prevailing wind being from the direction of the interior of the Yukon rather than the Pacific Ocean. It will, in general, be true that an air mass which has derived its water vapour from a continental rather than a marine source will precipitate water of a lower δ-value due to the lower δ-value of continental bodies of water. Such a phenomenon

could be true of an air mass moving from the direction of the interior of the Yukon as it did on the day of sample collection.

To support this idea of a secondary continental source of water vapour a number of δ-values for precipitation at Whitehorse, Yukon Territory, are shown in figure 19. These values have been measured by Dansgaard (personal communication). They show δ-values of summer precipitation of around -20 to -23. These would suggest that a value of $\delta = -22.6$ is indeed reasonable.

These samples of summer rainfall were used as a check on the internal consistancy of results. The first two samples were mixed at the time of collection so that they would represent two samples that should have the same δ-value. This turned out to be the case. The third sample had an initial composition identical to the first two. It was placed in a bottle that did not have an air tight cap with the idea of noting any evaporation by a change in the δ value of the sample. Evidently, no isotopic fractionations took place in excess of the $\pm$ 0.10 value quoted as the experimental error.

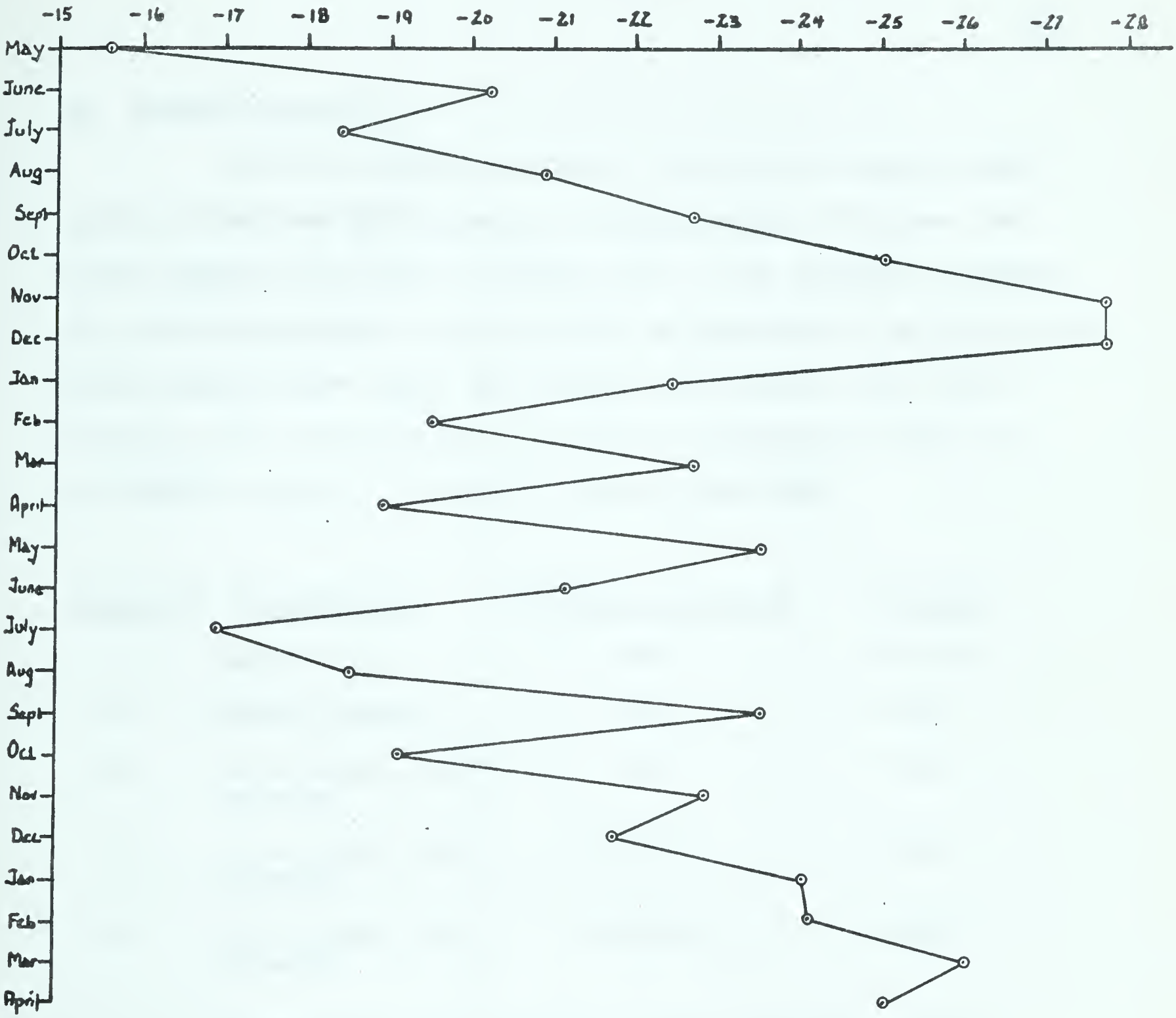


Figure 19

δ Values For Precipitation

at Whitehorse, Yukon

(Dansgaard)

d. Summer Snowfall.

After a summer snowfall, a series of samples were taken at various altitudes on a nearby peak. δ-values for these samples are shown in table VIII. The samples represent an attitude gradient of over 2000' as measured by an altimeter. Snow samples were taken by filling the aluminum tube with surface snow and heating the tube in a blowtorch flame in the usual fachion. Values for this study were:

Sample #	Description	Altitude above SL	δ value
1	moist firn	8203'	- 20.30
2	moist crusted	8525'	- 26.6
3	wind blown, moist, crusted	9647'	- 25.9
4	wind blown, dry, crusted	10,067'	- 20.9
5	wind blown, dry, crusted	10,510'	- 18.8

There does not seem to be any altitude effect present in these samples, nor do they show the degree of homogeneity that might be expected of snow samples from a single snowfall. The two values of δ = -25.9 probably represent winter snow. The spread of values for this study points to the pitfalls of sampling windblown surface snow that may have been transported from another snow accumulation site.

e. A borehole at the tope of the accumulation area.

On July 28, seven samples were taken from an auger core drilled through one annual layer. The sample site was located at an altitude of 8850 feet, which is close to the top of the accumulation area. Values for samples at this site are given in the table below.

Sample	Depth	δ Values
1	50 cms.	- 21.9
2	100 cms.	- 22.3
3	150 cms.	- 25.7
4	200 cms.	- 24.1
5	250 cms.	- 23.5
6	300 cms.	- 24.2
7	350 cms.	- 24.8

average δ = - 23.8

The previous year's summer surface was located at a depth of 360 cms. The average sample value for this pit is slightly more negative than the value for the first pit dug further down glacier. A more negative δ-value is predicted by the altitude effect. It is true that the least negative δ-values are found in snow near the surface of the annual accumulation which could be interpreted as being spring snow deposited at a warmer temperature than snow for the bulk of the winter. δ-values of -21.9 and -22.3 are close to the values obtained for the summer rainfall of July 7.

f. Pit number 2 on the Hubbard Glacier.

A second major pit study was carried out at a sample site located down the Hubbard Glacier. Pit number 2 was dug on July 17. Fifteen samples were taken with the idea of correlating their δ-values with stratigraphy and density measurements. Stratigraphy in this pit indicated that the melt season was well under way, as ice layers were evident throughout the whole stratigraphic column. The snow at this stage was isothermal to a depth of at least 310 cms. The previous summer layer was located at a depth of 243 cms. Evidently the winter accumulation was less at this site than sites on the Kaskawulsh side of the accumulation area. Apart from the volume of accumulation, the δ-values of snow samples from these two pits show considerable difference in their annual trend. The most negative δ-values in pit number 2 were found to be associated with snow much closer to the surface than in pit number 1. A marked discordance occurs between samples, above and below the fall-winter boundary in pit number 2, but in this case, the snow above this layer was less negative than snow below it, contrary to what would be expected. Figure 20 shows that the density of snow samples did not decrease below the fall-winter interface as they did in pit number 1 (figure 17). It is interesting to note (in this instance) the correlation between the oxygen isotopes and density measurements. Figure 20

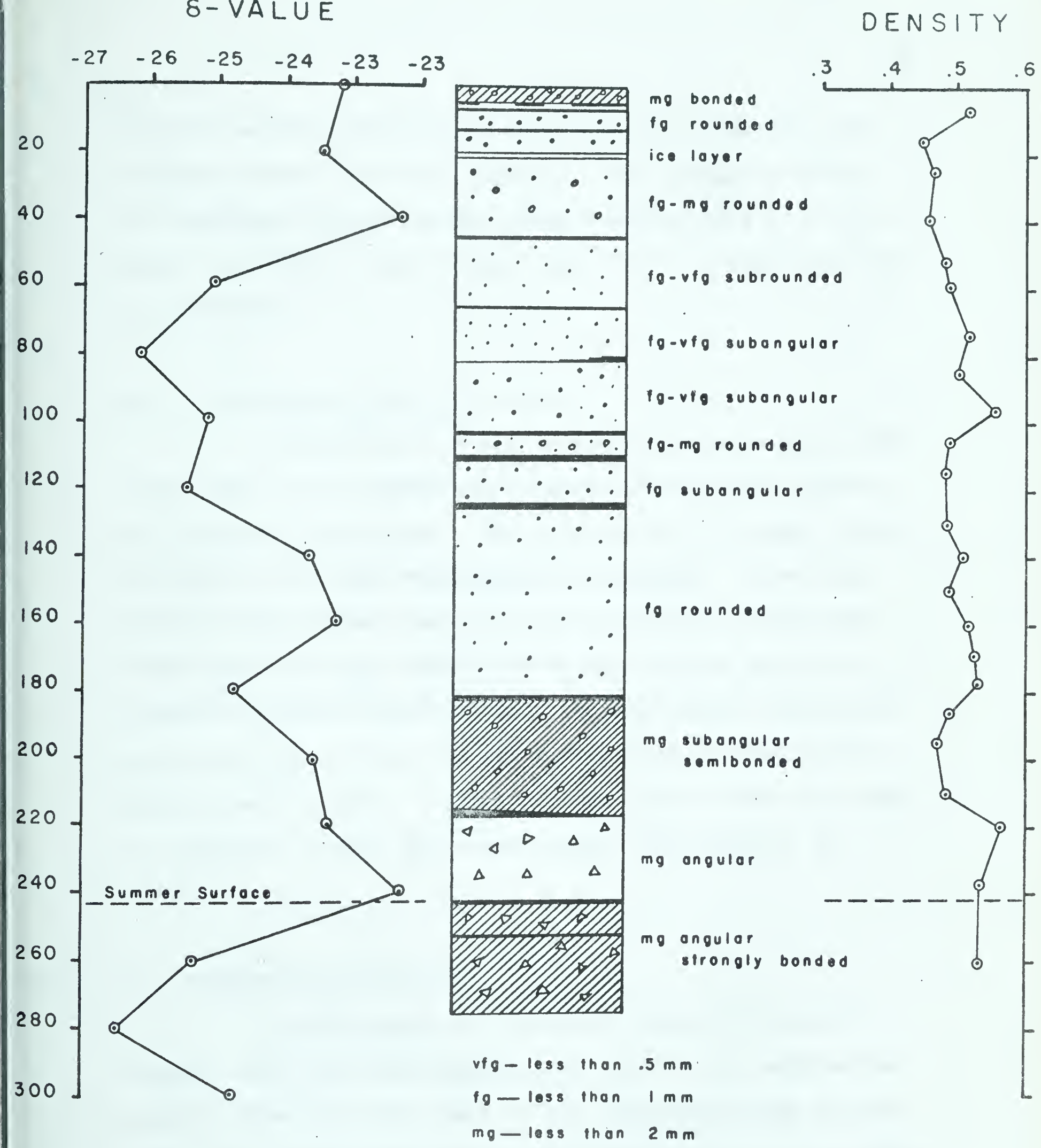


FIGURE 20

PIT NUMBER 2

Showing the Relationship Between Oxygen Isotopes, Stratigraphy, and Density

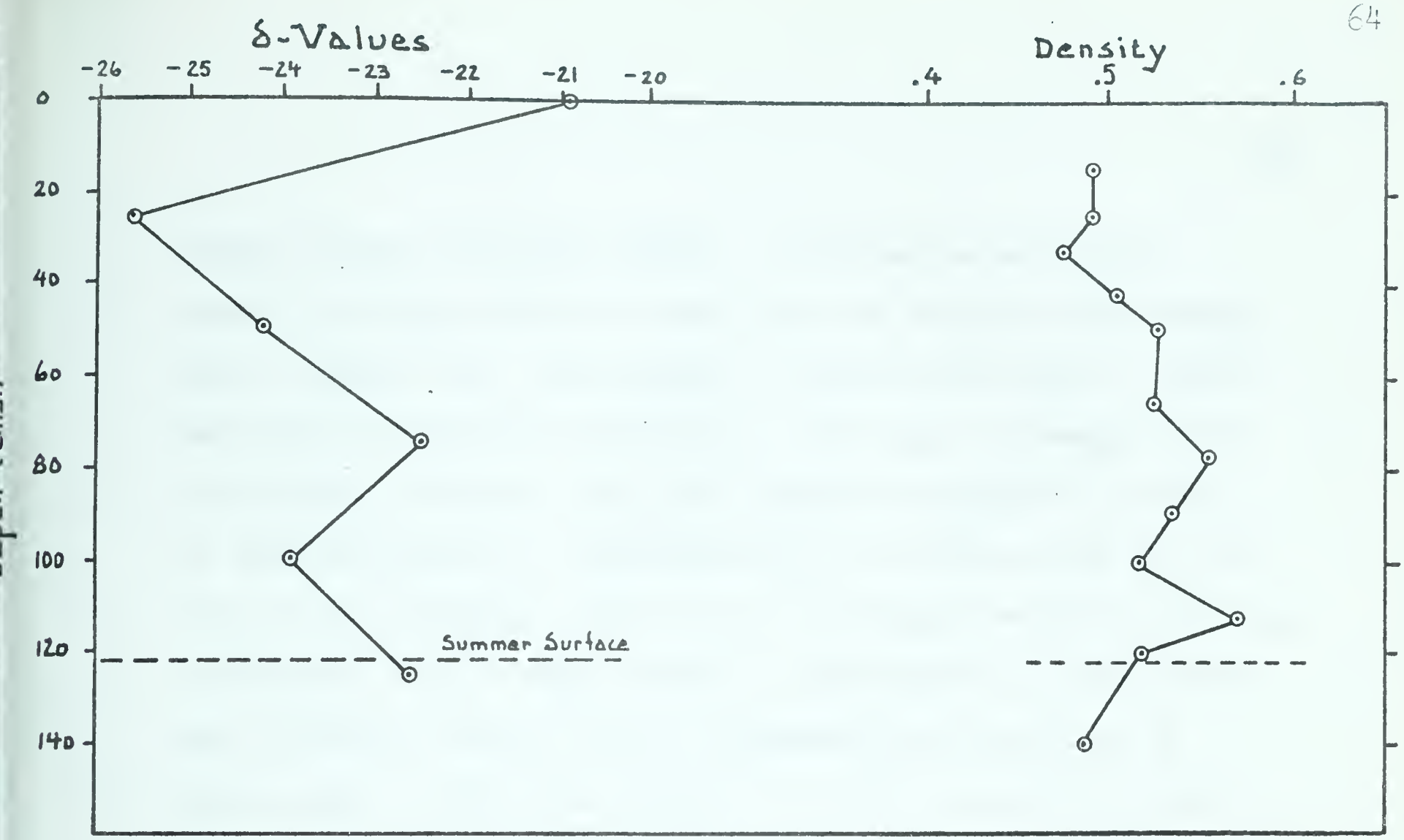
also illustrates the annual trend that is present in the oxygen isotopes from pit number 2. The average δ-value for samples from pit number 2 was found to be δ = - 24.0, which is close to the average value of δ = - 23.1 found for pit number 1.

g. A pit study at the firn line.

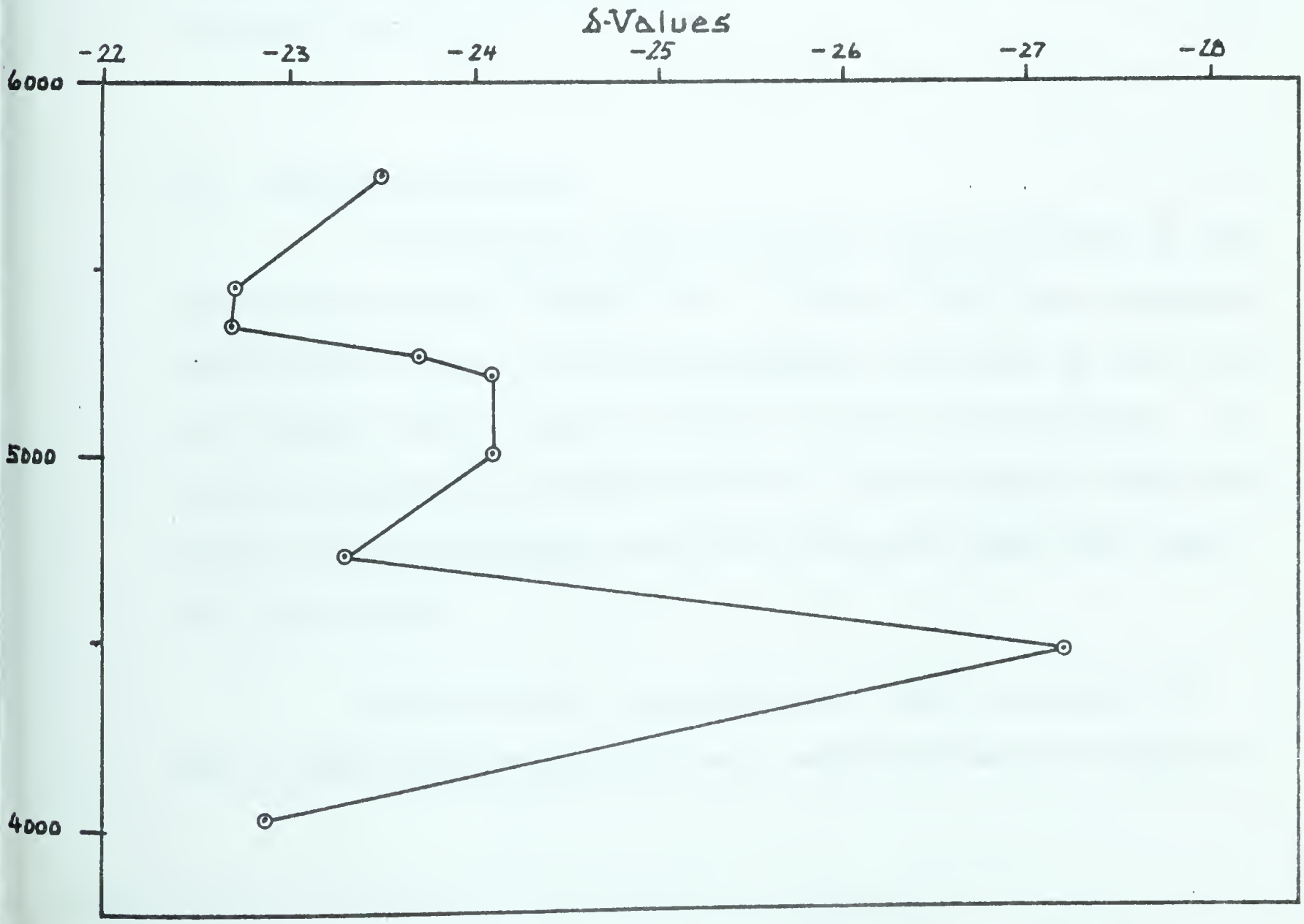
A third pit was dug on August 22 as close to the firn limit on the Kaskawulsh glacier as the water content of the snow would allow. The location of pit number three is shown on the map reproduced in figure 21. Snow from this pit was almost completely saturated with water from summer melt action, making the oxygen isotope ratios of somewhat limited significance. The fall-winter interface was located at a depth of 122 cms. by density and stratigraphy measurements. Values for pit number three are taken to represent little more than average for snowfall at that site.

h. Longitudinal Profile.

Shortly after the furthest extent of the melt season, nine hard ice samples were taken on a longitudinal profile from the firn limit to the terminus of the glacier, with the idea of detecting an altitude effect in the oxygen isotopes. Samples were spaced 2 miles apart, and in each case,



Pit Number 3 Figure 21



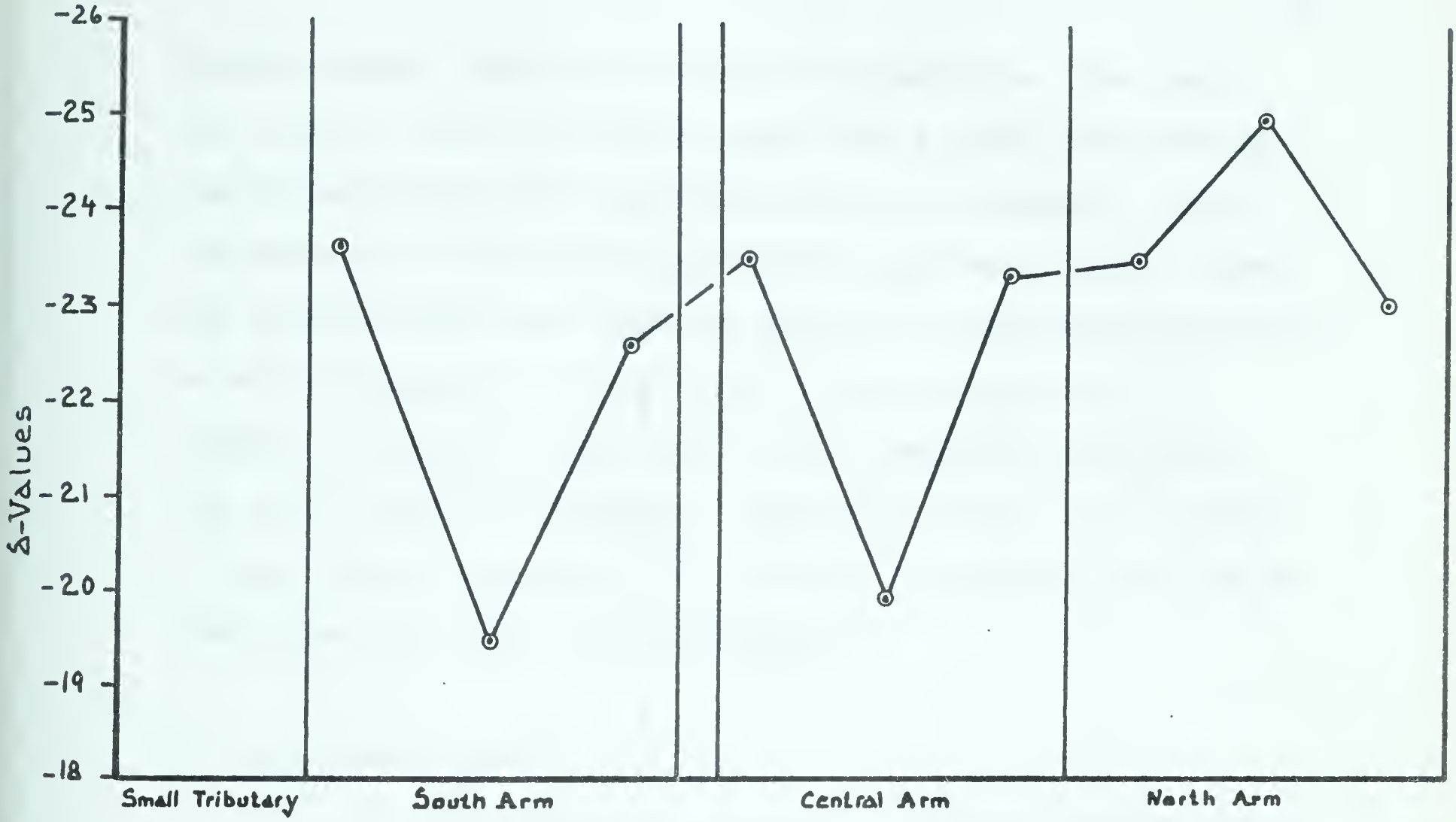
Longitudinal Profile Figure 22

coarse, bubbly ice was sampled. Altitude measurements quoted in figure 22 were taken from the Dominion government map in figure 16. The scatter of results show that a great many more samples are required to make any noticeable trend conclusive. However, the nine samples do suggest a trend, in that the value of δ decreases with altitude from the firn line to the terminus. This is in accordance with the finding of Epstein for the Blue Glacier in Washington. Any noticeable altitude effect for the Kaskawulsh glacier will be complicated by the flat geometry of the accumulation basin which would reduce any noticeable altitude effect. The complex meteorology of the area will present a further complication.

i. Transverse Profile.

A transverse sample profile also was taken on the Kaskawulsh glacier (figure 23). The site for the transverse profile was chosen to include samples from each of the three ice streams which make up the main body of the glacier. As in the longitudinal sample profile, the extremely large size of the glacier warranted many more samples than were taken for this study.

It was found, nevertheless, that the center of two of the ice streams were less negative than the edges of



Transverse Glacier Section

Figure 23

these streams. This would tend to support the idea that the center of an ice stream comes from a lower altitude in the accumulation area than the sides. As a result, ice in the center of the stream has a less negative δ-value. Similar results have been found by Epstein on transverse profiles of other glaciers. Of some note is the consistency of values for samples taken next to the moraines separating the three major ice streams. These values are close enough to each other that a similar altitude of accumulation can be postulated for each of these samples.

j. Ice-cored moraine.

On the Kaskawulsh glacier, a crevass running from one ice stream through to the next ice stream exposed the ice core of a medial moraine. The sample site was on the moraine between the north and central ice streams parallel to the site of sample number 4 on the longitudinal profile (figure 16). The sample was analyzed and found to have a δ-value of - 22.7; not significantly different from the average of the hard ice samples. It is difficult to postulate an origin for this sample other than formation from ice derived from one of the main streams.

k. Meltwater from the Terminal Moraine.

A composite sample was made from meltwater streams emerging from the terminal moraine. The δ-value of this sample was - 22.6; a value very close to the average of the hard ice samples and the ice core of the medial moraine.

δ-values for ice samples taken below the firn line were found to have a smaller range than snow samples taken from pits in the accumulation area. With very few exceptions the ice samples closely represent the mean value of all the samples taken (δ = - 22.8). The average value of each of the pit studies was found to be quite close to δ = - 22.8. Such a consistency of results suggests that the sampling has been consistent as well.

The scatter of experimental results in this and similar oxygen-isotope studies point to the need for more extensive meteorological control for each of the ice and snow samples.

If a year-round meteorological station were established on the glacier, correlation of oxygen-isotope studies would be greatly facilitated.

APPENDIX

Samples from Pit Number 1

Sample Number	Depth (cms.)	δ Value
1	0	- 18.2
2	20	- 17.1
3	40	- 18.3
4	60	
5	80	
6	100	- 22.3
7	120	- 20.9
8	140	- 23.1
9	160	- 23.2
10	180	- 23.4
11	200	- 26.3
12	220	- 26.7
13	240	- 29.3
14	260	- 24.9
15	280	- 23.2
16	300	- 23.4
17	320	- 23.3
18	340	- 23.9
19	360	- 25.2
20	380	- 22.3
21	400	- 22.3
22	420	- 22.5

average value = - 23.1

of samples above 360 cms.

Table I

Table II

Ice Layers from Pit number 1

Sample Number	Depth (cms)	δ-Value
1	20	- 22.5
2	31	- 20.9
3	39	- 23.4

Table III

Snow Samples from Pit Number 2

1	0	- 23.2
2	20	- 23.5
3	40	- 22.3
4	60	- 25.1
5	80	- 26.2
6	100	- 25.2
7	120	- 25.5
8	140	- 23.7
9	160	- 23.3
10	180	- 24.8
11	200	- 23.6
12	220	- 23.4
13	240	- 22.3
14	260	- 25.4
15	280	- 26.5
16	300	- 24.8

average value = - 24.0

of samples above 240 cms.

Table IV

Ice Samples from Longitudinal Profile

Sample Number	Altitude	δ Value
1	5750'	- 23.5
2	5450'	- 22.7
3	5350'	- 22.7
4	5270'	- 23.7
5	5210'	- 24.1
6	5000'	- 24.1
7	4730'	- 23.3
8	4470'	- 27.2
9	4329'	
10	4020'	- 22.9

average value of
ice samples = - 23.8

Table V

Ice Samples from Transverse Profile

Sample Number	δ Value
1	- 23.0
2	- 25.0
3	- 23.5
4	- 23.3
5	- 20.0
6	- 23.5
7	- 22.6
8	- 19.5
9	- 23.6

Average value of
ice samples = - 22.7

Table VI

Snow Samples From Pit Number 3

Sample Number	Depth (cms)	δ Value
1	0	- 20.9
2	25	- 25.6
3	50	- 24.2
4	75	- 22.5
5	100	- 23.9
6	125	- 22.6

Average Value = - 23.3

Table VII

Samples from 7 meter hole at the site of Pit number 1

Sample Number	Depth (cms)	δ Value
1	0	- 19.7
2	50	- 21.8
3	100	- 19.7
4	200	- 24.3
5	250	- 21.9
6	300	- 19.6
7	350	- 24.7
8	400	- 25.7
9	600	- 19.7
10	700	- 22.6

Average Value = - 22.3

Table VIII

Samples of Summer Snowfall on a Nearby Peak

Sample Number	Altitude	δ Value
1	8203'	- 20.3
2	8525'	- 26.6
3	9647'	- 25.9
4	10,067'	- 20.9
5	10,510'	- 18.8

Table IX

Samples Taken from an Auger Hole at the Top of the Accumulation Area

Sample Number	Depth (cms)	δ-Value
1	50	
2	100	- 22.3
3	150	- 25.7
4	200	- 24.1
5	250	- 23.5
6	300	- 24.2
7	350	- 24.81

Average Value = - 24.1

Table X

Summer Rainfall Sample

Sample Number	δ Value
1	- 22.6
2	- 22.6
3	- 22.5

Table XI

Sample from Ice-Cored Moraine

Altitude	δ Value
5270'	- 22.7

Table XII

Samples from Terminal Moraine Water Run-Off

Altitude	δ Value
7000'	- 22.6

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